International Geology Review

PARTIAL CONTENTS

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International Geology Review is published monthly by the American Geological Institute with the assistance of an initiating grant from the National Science Foundation. The journal will report, in English, significant contributions to pure and applied research in the earth sciences which appear in foreign-language journals, especially those published in the U.S.S.R.

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International Geology Review

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REFERENCE SECTION

IGR TRANSLITERATION OF RUSSIAN

The AGI Translation Office has adopted the Cyrillic transliteration recommended by the U. S. Department of the Interior, Board on Geographic Names, Washington, D. C.

NOTES:

(1) "ye" initially, after vowels, and after ъ, ь ; "e" elsewhere; when written as "ё" in Russian, transliterate as "yё" or "ё".

Alphabet		transliteration
А	а	a
Б	б	b
В	в	v
Г	г	g
Д	д	d
Е	е	e, ye [1]
Ё	ё	ё, yё
Ж	ж	zh
З	з	z
И	и	i
Й	й	y
К	к	k
Л	л	l
М	м	m
Н	н	n
О	о	o
П	п	p
Р	р	r
С	с	s
Т	т	t
У	у	u
Ф	ф	f
Х	х	kh
Ц	ц	ts
Ч	ч	ch
Ш	ш	sh
Щ	щ	shch
Ъ	ъ	"
Ы	ы	y
Ь	ь	'
Э	э	e
Ю	ю	yu
Я	я	ya

Well-known place and personal names that have wide acceptance will be used. Some translations may include elements of previous German transliteration from the Russian; this occurs in IGR most commonly in maps and lists of references. The reader's attention is called to the following variations between German and English systems which may cause confusion when trying to check back to original Russian sources.

German	English
w	v
s	z
ch	kh
tz	ts
tsch	ch
sch	sh
schtsch	shch
ja	ya
ju	yu

THE GENESIS OF KYANITE IN QUARTZ VEINS[1]

by

Ivan Kostov[2]

• translated by Royer and Roger, Inc. •

ABSTRACT

The article deals with a new vein deposit of kyanite in Bulgaria and some considerations concerning the origin of kyanite veins in general. The kyanite is found in quartz veins within garnetiferous staurolite schists of probable Jurassic age. The kyanite crystals are up to ten centimeters long along the c-axis and form parallel rodlike aggregates perpendicular to the selvages of the veins. The origin of the kyanite in the quartz veins is due to mobilization of silicon and aluminum under the action of boron and small quantities of alkalies, contained in pneumatolitic-hydrothermal or hydrothermal solutions. The temperature and the pressure are thought rather low but should be taken into consideration. It is suggested that the terms "stress" and "antistress minerals" be dropped and the terms "piezogene" and "thermogene" be introduced, depending on whether pressure or temperature plays the prevalent role in the formation of the mineral. -- auth. English summ.

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Of the three forms of $Al_2O(SiO_4)$, kyanite is perhaps the most interesting both mineralogically and crystallochemically. Kyanite is usually considered to be an indication of very great pressure, and has been placed by A. Harker (1939) in the first rank of the group of so-called stress-minerals. But kyanite occurs in nature in two forms: a) disseminated in gneisses and mica schists, and b) in pegmatites, quartz veins and rock crystal. The second group is almost always continuously associated with the first. Consequently one may accept the origin attributed to kyanite deposits by Harker in the case of the first group. In the case of the second type, however, such an origin is illogical and hard to believe, in view of the fact that the kyanite crystals are freely formed, as has been established, in druse cavities within the quartz veins (Miyashiro, 1951).

Some data will be given below on the kyanite in a new vein deposit in Bulgaria and generalizations will be made from observations of the origin of this type of deposit.

CONDITIONS OF OCCURRENCE OF KYANITE

The kyanite described in this article was found north of the village of Dervishka Mogila, in Mt. Sakár in Eastern Bulgaria; it was located in quartz veins up to 20 cm thick parallel to the foliation of a garnet-staurolite schist that did not contain any kyanite. The schist, extending from east to west and dipping steeply (about

60°) southward alternates with interbeds of amphibolite. The garnet-staurolite schists consist of large almandite and staurolite porphyroblasts, as well as flakes of biotite and muscovite streaked with quartz (fig. 1). The garnet porphyroblasts form euhedral rhombic dodecahedra up to 1 cm in diameter, some of them with poorly developed (211) faces. These porphyroblasts are generally homogeneous, but frequently interpenetrated with quartz, in distinction to the garnet porphyroblasts of similar schists occurring about 1 or 2 km south of the kyanite veins; here the garnets are zonal and contain inclusions of quartz (fig. 2). These facts indicate recrystallization of the garnets of the first location, or a lesser degree of metamorphism of the garnets of the second type, whose paragenetic association does not include staurolite. The staurolite from the garnet-staurolite schists has prisms along (110), with the (010) and sometimes (101) faces well developed. The crystals vary up to 3 cm in length. The biotite is fresh, and shows brown and orange pleochroism. Muscovite occurs as small flakes. Acidic plagioclase (oligoclase) is found, but quite rarely. The quartz is characterized by an even extinction grading into slightly wave-like extinction, indicating some small degree of deformation after crystallization. Although the quartz occurs as inclusions in the garnet and staurolite porphyroblasts, and also in the biotite laminae, the sporadically encountered interlayed veinlets of the same quartz in the staurolite crystal units indicates that it was the latest mineral to be formed. The order of formation of the prophyroblastic minerals is as follows:

$$garnet \rightarrow staurolite \rightarrow biotite \rightarrow muscovite \rightarrow quartz.$$

Among the accessory minerals in the garnet-staurolite schists, tourmaline is abundantly developed in the form of small olive-green crystals. There is also a radioactive mineral that
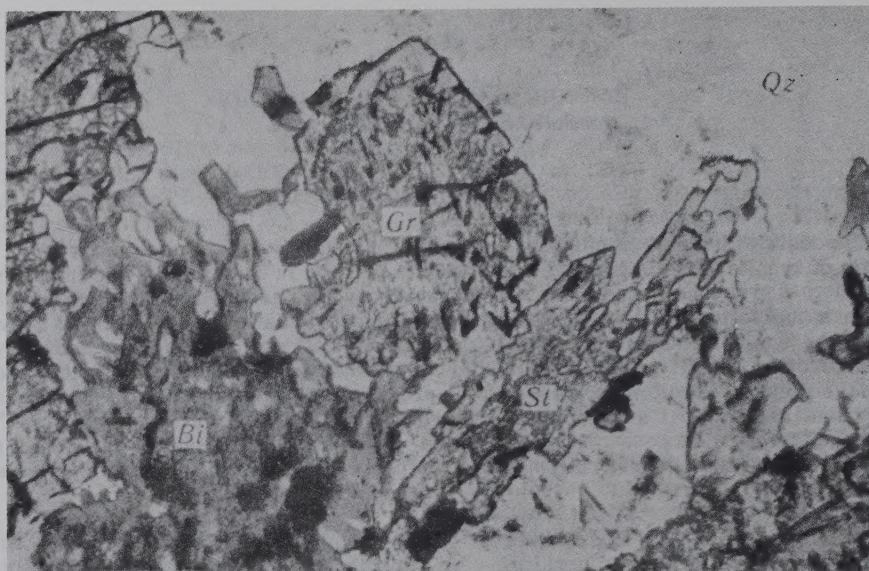
FIGURE 1. Garnet-staurolite schist: Gr - garnet, st - staurolite, Bi - biotite, Qz - quartz; magnification x 50.
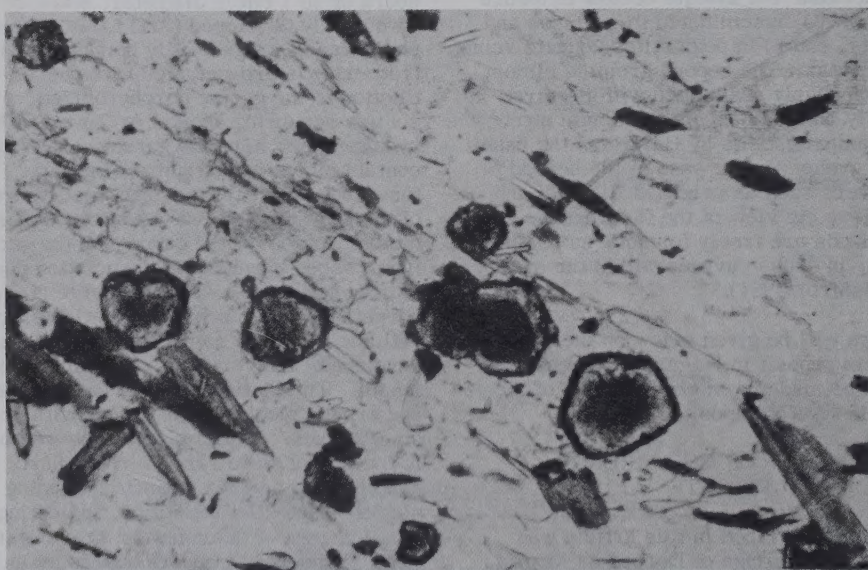
FIGURE 2. Garnet schist. Magnification x 50.

forms crystals of longitudinal prismatic habit, high birefringence, refractive indices like those of apatite and negative elongation. Sometimes one may observe pyrite and oval grains of apatite. More apatite, along with rutile, occurs in the amphibolites, consisting mainly of blue-green amphibole with an absorption pattern $\beta > \epsilon > \omega$ and $c\epsilon = 18°$, as well as quartz with an even or slightly wave-like extinction. The amphibole crystals are acicular prismatic,

bounded by $\{110\}$ faces, with $\{100\}$ well developed and $\{010\}$ poorly represented. Sometimes the following minerals are also encountered: clinozoisite, magnetite and, in the amphibolites garnet and anorthite.

Apart from the quartz-kyanite veins, the garnet-staurolite schists and amphibolites also contain quartz veins with tourmaline (schorlite) and more rarely with micaceous

matite or epidote. Some of the tourmaline
[gra]ins have a pegmatoid character and contain,
[in] addition to the quartz and tourmaline, an
[aci]dic plagioclase with refractive index
[n ≈] 1.54 strongly sericitized, with a vaguely
[lam]inar texture that tends toward spottiness.
[The] tourmaline crystals have no crowning faces,
[the] prismatic zone is bounded by {1120} and
[{1]10}, and the individual specimens usually
[hav]e the appearance of cluster-like to radiating
[ag]gregates. The tourmaline is distinguished
[by] its clear zonal structure, consisting of a
[blu]e or blue-green core and a brownish outer
[zo]ne.

[North of the kyanite veins the garnet-stauro-
[lit]e schists and amphibolites merge with a
[blu]ish-gray and white marble next to a porphy-
[rit]ic microcline granite that forms the higher
[pa]rts of the mountain and lies directly adjacent
[to an] injection gneiss. Younger than the strati-
[fie]d complexes are the dike rocks south and
[ea]st of Topolov-grada, in the northern part of
[Mt.] Sakar.

[The time of formation of the schists that
[co]ntain the kyanite veins is still unknown. They
[ha]ve been called Jurassic (Yanishevski, 1946;
[Dim]itrov, 1955). The quartz-kyanite veins
[we]re formed after the metamorphism of the
[Ju]rassic sedimentary series and before the
[fo]rmation of the polymetallic and barite-hema-
[tit]e veins that occur to the northeast of the
[gra]nite deposits.

PROPERTIES OF THE KYANITE

The kyanite occurs exclusively in quartz

Chemical compositions of dark blue kyanite, light
kyanite and muscovite formed by the alteration of
kyanite

Components	Dark blue kyanite	Light kyanite	Muscovite
SiO_2	36.11	33.73	43.11
TiO_2	-	-	0.16
Al_2O_3	59.80	62.10	38.15
Fe_2O_3	2.96	1.10	2.47
FeO	0.18	0.36	0.18
MgO	0.27	0.53	0.50
CaO	0.53	0.60	0.54
Na_2O	0.55	1.19	1.98
K_2O	0.28	0.73	5.55
H_2O+	-	0.82	5.48
H_2O-	-	0.12	2.48
	100.68	100.89	100.60

veins as parallel columnar aggregates, the
individual units of which are highly elongated
along the c axis and almost directly perpendicu-
lar to the vein walls. The kyanite crystals vary
in length up to 10 cm in the direction of the c
axis. The crystals are depressed along the
first pinacoid {100} (figs. 3, 4). In addition to
{100}, {010} and {001} and, in some crystals,
the (110) faces are poorly developed. In view
of the known post-crystallization movements
occurring in certain veins, the kyanite crystals
are slightly deformed and displaced from their
original positions. The kyanite is blue in color,
grading to light blue in the axial part of the crys-
tals, and along the outer margins paling to al-
most white; this is associated with the partial
change in its composition or the transition from
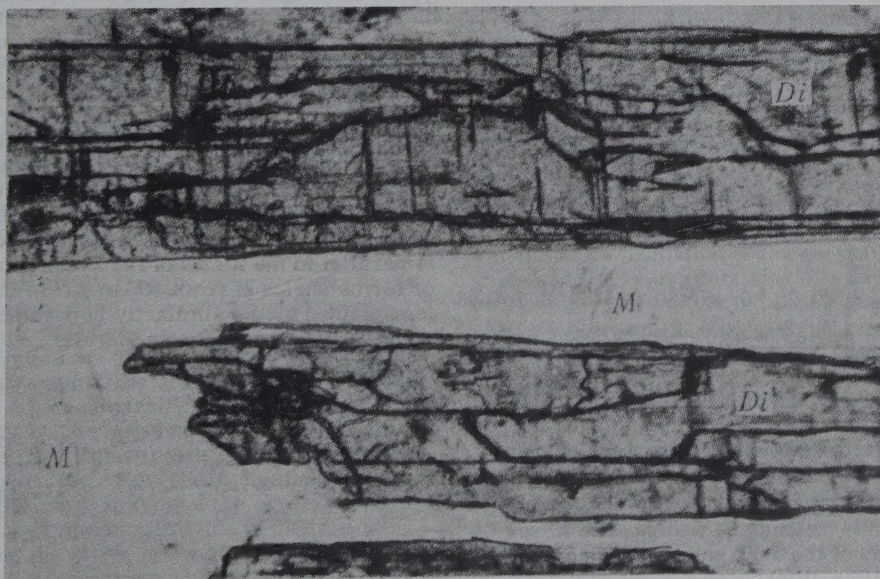
FIGURE 3. Kyanite (Di) and muscovite (M). Magnification x 50.
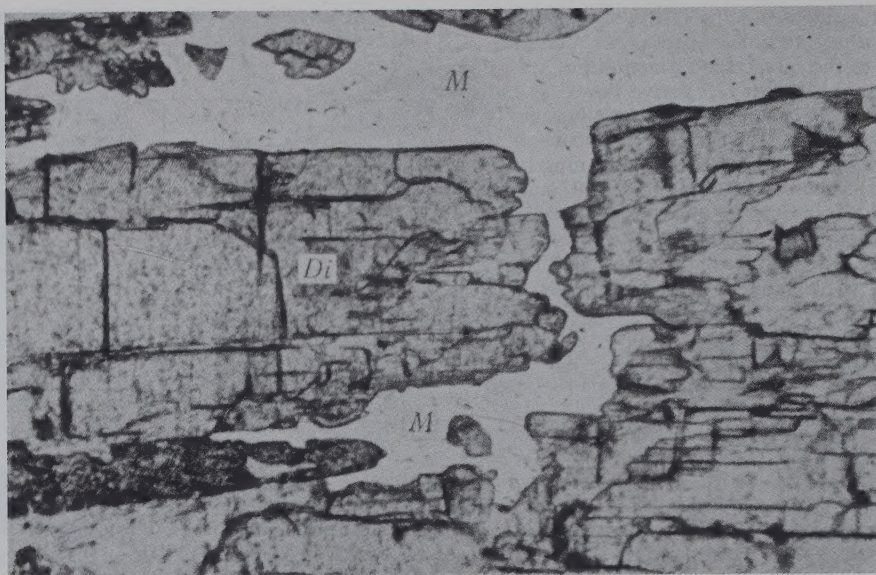
FIGURE 4. Kyanite (Di) and muscovite (M) formed in fractures. Magnification x 50.

the kyanite to muscovite. Pleochrosim: ϵ — azure blue, β — light blue to colorless, ω — colorless; $c \epsilon = 30 - 32^o$. In the highly colored blue kyanites $2V = 84^o$; in the almost colorless and white forms $2V = 86^o$. The table below shows the chemical composition of the dark blue and light-colored kyanites along with that of the muscovites formed by alteration of the kyanites.

Recomputation of the results of this analysis resulted in the following crystallochemical formulas:

Dark blue kyanite - $(Na_{0.121} K_{0.04} Ca_{0.06})_{0.22}$

$(Al_{7.57} Fe^{3+}_{0.23} Fe^{2+}_{0.03} Mg_{0.06})_{7.96}$

$[(Si_{3.93} Al_{0.07}) O_{20}]$

Light kyanite - $(Na_{0.24} K_{0.09} Ca_{0.07})_{0.40}$

$(Al_{7.71} Fe^{3+}_{0.09} Fe^{2+}_{0.03} Mg_{0.08})_{7.91}$

$(Si_{3.87} Al_{0.13}) O_{20}]$

Muscovite - $[Na_{0.25} K_{0.46} Ca_{0.03} (H_3O)_{0.24}]_{0.98}$

$(Al_{1.87} Fe^{3+}_{0.13} Fe^{2+}_{0.01} Mg_{0.05})_{2.06}$

$[(Si_{2.88} Al_{1.12}) O_{10}] (OH)_2 \cdot 0.46 H_2O.$
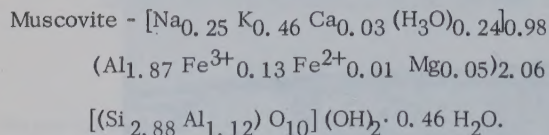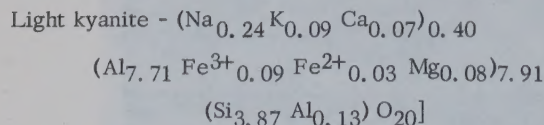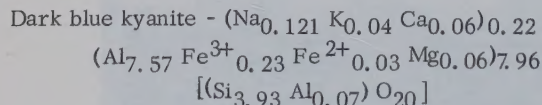
The alterations of the kyanite were evidently accompanied by a decrease in the amounts of Fe^{3+} and Si^{4+} and an increase in the content of Al^{3+}, alkaline cations and water. The X-ray micrographs of the dark blue and light kyanites show no visible differences.

The change from kyanite to muscovite is

accompanied by a sharp increase in alkaline elements, particularly potassium, with a considerable decrease in the aluminum in the composition of the kyanite. The muscovitization of the kyanite crystals was in all probability due to high-temperature aqueous solutions saturated with alkalines, which extracted mainly silica and aluminum. This process, it must be assumed, was associated with the formation of laumontite and chalcedony in fractures in the quartz-kyanite and certain pegmatoid quartz veins. In the case of the mica, $\epsilon = 1.590$, $\beta = 1.588$ and $2V = 37^o$; this is characteristic of partially hydrated muscovite. Although the kyanite crystals contain irregular fractures perpendicular to their long dimension (fig. 4), the muscovitization is nevertheless observed either within these or else along the surfaces of the usual fractures along (100). In the first case the muscovite flakes are oriented at random relative to each other, whereas in the second case the replacement is of an epitaxitic nature: the a axis (ϵ) of the muscovite crystals is parallel to the a axis of the kyanite crystals or forms angles of from 30° to 90° to it. This is explained by the similarity in parameters of the structures of the two minerals:

percent

Kyanite.....	$c_o = 5.56 \overset{o}{A}$	$\Delta = 7.3$
Muscovite ..	$a_o = 5.18$	
Kyanite.....	$b_o = 7.72$	$\Delta = 14.1$
Muscovite ..	$b_o = 9.02$	

The difference in either case does not exceed the limits imposed by epitaxitic growth.

ORIGIN OF THE KYANITE

In studying the genesis of the kyanite deposit described here, one must keep in mind the following circumstances: 1) the kyanite deposits are conformable to the layering of the garnet-staurolite schists, and the crystals of kyanite are highly elongated; 2) the garnet-staurolite schists themselves do not contain kyanite, but do contain a large quantity of tourmaline concentrated in quartz veins like the kyanite; 3) the kyanite crystals, elongated along the c axes, are arranged as aggregates of parallel columns perpendicular to the vein walls; 4) the kyanite veins occur in areas characterized by a lamellar texture and folded structures complicated by numerous intrusive bodies.

Staurolite, according to Harker (1939), is a typical stress mineral stable within a small temperature interval. It is transformed into kyanite plus almandite garnet according to the following equation:

$$3FeAl_4Si_2O_{10}(OH)_2 + 2SiO_2 = 5Al_2SiO_{15} +$$
$$\text{staurolite} \quad \text{quartz} \quad \text{kyanite}$$

$$+ Fe_3Al_2Si_3O_{12} + 3H_2O$$
$$\text{almandite}$$
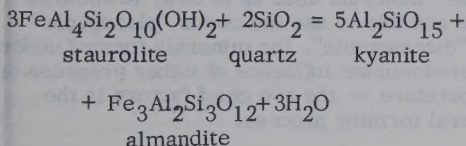
The associations of staurolite plus almandite (as observed in this case) and staurolite plus kyanite are frequently encountered as transitional types between garnet schists, on the one hand, and garnet-kyanite schists, on the other (Harker, 1939; Jagitsch, 1949; Semenenko, 1953).

The absence of kyanite in the garnet-staurolite schists of the Sakar deposit indicates sharply differing conditions of formation of the kyanite in this case. Since the crystals of this mineral are perpendicular to the quartz veins, they have evidently grown freely in open fractures in the garnet-staurolite schists. This is also shown by the presence of quartz, which is of later formation than the kyanite and whose c axes are also predominantly perpendicular to the walls of the veins.

It is well known that the structure of kyanite may be represented as a cubic closest packing of oxygen ions (Miyashiro, 1951) in which all the aluminum ions are surrounded by octahedra of oxygen. Half of the aluminum ions are connected in columns parallel to the c axis and constructed in such a manner that the layers of AlO_6 octahedra correspond to the planes of the first pinacoid (fig. 5). These aluminum octahedra apparently determine the crystal habit of the kyanite crystals, while the remaining aluminum and the silica cations separate the layers of octahedra from those lacking aluminum octahedra, thus causing the perfect cleavage of kyanite along (100). The orientation of the
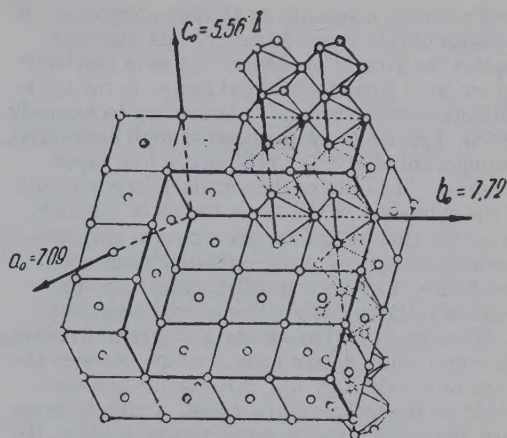
FIGURE 5. Diagram of the structure of kyanite, showing the closest-packing arrangements of the oxygen atoms and the positions of the columns of aluminum-oxygen octahedrons.

kyanite porphyroblasts along (100) relative to the layering of the surrounding rocks corresponds closely to this structural peculiarity of the mineral, and to Rikke's principle of crystallization under the conditions of directional pressure. In deposits of the vein type, and in the Mt. Sakar deposits under consideration here, the kyanite crystals have their long dimensions (corresponding to the direction of the aluminum-oxide columns of octahedra along the c axes) perpendicular to the walls of the quartz veins. Hence one cannot assume a pressure acting in this direction. In this case the kyanite crystals grew in a free and open medium without the hindrance of directional pressure, which evidently played a great role in the formation of the garnet-staurolite schist layers. Thus a definite analogy may be drawn between the kyanites of the two types of deposits and the almandine-garnet schists and the corresponding pegmatite veins. The kyanite crystals in the deposits of the vein type contain no quartz inclusions, in distinction to the kyanite of the kyanite schists, which is interpenetrated with quartz like the garnet and staurolite of the deposit considered here (figs. 1, 3, 4). In view of the simultaneous formation of the quartz-kyanite veins and the veins containing other minerals, particularly tourmaline, which is abundantly scattered throughout the garnet-staurolite schists, pneumatolytic-hydrothermal action must be considered to be their most likely origin. In this case there will have been a removal of silica and aluminum and of part of the alkali elements. This resulted in the creation of an aluminum concentration with the formation of kyanite or tourmaline, arising when there was a considerable content of iron (Michel-Levy, 1950), probably from the staurolite or the biotite in the schists.

Kyanite in vein deposits must not be consid-

ered a stress mineral, as Harker supposes. In deposits of this type, however, one must not neglect the role of pressure, not as a mechanical but as a physicochemical factor in the crystallization of the kyanite. According to Kennedy (1955), kyanite is synthesized at high pressures, a temperature of 550 - 600° and a low vapor pressure. In spite of the repeated experiments in the synthesis of the three forms of Al_2SiO_5, up to this time there are no accurate physicochemical diagrams of the stability ranges of andalusite, kyanite and sillimanite. On the basis of microscope studies of thin sections, A. Hietanen (1954) proposes a diagram stressing the empirically known relationships between the three minerals (fig. 6). Sillimanite becomes stable as the temperature rises, a rise in pressure and a decrease in temperature leads to the formation of kyanite, and both minerals become andalusite by metasomatism under static conditions.
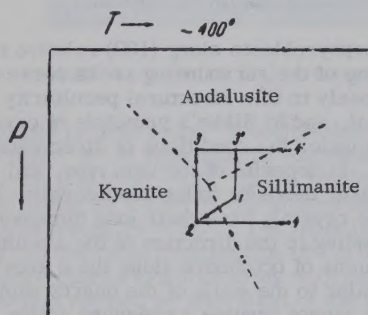
FIGURE 6. Hypothetical diagram of the fields of stability and alteration of the three forms of Al_2SiO_5. The figures show the respective transitions from the first to the second forms (after A. Hietanen, 1954).

The genesis of kyanite at high pressure is also indicated by its crystallochemical properties, considering the fact that an increase in the pressure should facilitate physicochemical reactions with a higher coordination number — that is, as usual it should have an effect opposite to that of raising the temperature (Sobolev, 1949, 1955). According to the equilibrium diagram shown in fig. 6, kyanite is the lowest-temperature mineral of the three forms of Al_2SiO_5. V. S. Sobolev came to the same conclusion in studying the thermal effects of kyanite and andalusite (1956). The relatively low temperature of formation of the kyanite in the quartz veins of Mt. Sakar is also indicated by the minerals in the other quartz veins, particularly the epidote and the micaceous hematite.

A. N. Igumnov and K. Ye. Kozhevnikov (1935) assert that the Urals kyanite deposits, or rather the kyanite in both the veins and the

schists of the Borisov deposit, was formed hydrothermally by the action of the residual products of the magma. This may also be true of the formation of diabrochite (Ishioka and Suwa, 1956) considering that kyanite schists are believed by N. P. Semenenko (1954) to be the products of so-called migrational contact metamorphism.

Generalizing from the above, it may be said that the genesis of the kyanite in the quartz veins is definitely associated with the removal of silica and alumina by the boron and a small amount of alkali contained in pneumatolytic-hydrothermal or only hydrothermal solutions at comparatively low temperatures and small pressures. In the layered deposits the kyanite was formed at higher temperatures and under relatively high pressures, according to the diagram suggested by Hietanen (1954), although the significance of pneumatolytic-hydrothermal agents must not be neglected in this case either. Thus instead of the terms "stress" and "antistress" minerals used up to now, it would be more suitable to use the terms "piezogenic" and "thermogenic", for minerals formed under the predominant influence of either pressure or temperature — the two chief factors in the mineral forming process.

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ON THE PROBLEM OF THE GENESIS
OF CINNABAR AND METACINNABARITE [1]

by

G. L. Piotrovskiy [2]

•   translated by Royer and Roger, Inc.   •

ABSTRACT

Experiments have been made on the synthesis of cinnabar and metacinnabarite in conditions close to those of hydrothermal processes of the formation of mercury deposits in nature. If may be concluded that during the formation processes of mercury the first to fall out is metacinnabarite. The fate of metacinnabarite depends on the medium that surrounds it. If hydrothermal waters carrying alkaline solutions Me_2HgS_2 will continue to effect the formed metacinnabrite, the latter will gradually become a more stable red modification HgS, i.e., cinnabar. If the inflow of the mercury sulfide solutions stops or becomes so weak that under the counteraction of acid or even neutral vadose waters there will not be enough solvent in the form of ions S^{2-}, necessary for the changing of metacinnabarite into cinnabar, then the former will be preserved in the same state for a long time. --auth. English summ.

* * *

There are still many gaps in our knowledge of the genesis and interrelationships of cinnabar and metacinnabarite.

According to the commonly accepted theory of the origin of mercury deposits, cinnabar is precipitated from alkaline hydrothermal sulfide solutions. Metacinnabarite usually occurs in the oxide zone of many mercury deposits, and is thus generally considered to be a secondary, exogenic mineral formed as a result of the action of acidic surface solutions. Metacinnabarite is very rarely mentioned in the literature as a primary hydrothermal mineral.

Recently a number of papers by authors who have studied the mercury ore occurrences in Transcarpathia, on the basis of analyses of the paragenetic relationships of the minerals in these ore formations, have indicated that in many places in Transcarpathia the metacinnabarite is a primary hydrothermal mineral, and cinnabar is the secondary mineral developed after metacinnabarite (Barishnikov et al., 1957; Lazarenko, 1957; Merlich, 1957).

It is to be expected that experiments in synthesizing metacinnabarite in the laboratory would be of great help in deciding whether this mineral is primary or secondary. Unfortunately, however, the existing data on the synthesis of metacinnabarite are far from sufficient to answer this question. The small number of accounts describing the synthesis of metacinnabarite date mainly from the end of the last and

beginning of this century. The defect of these experiments lies in the fact that the conditions under which the metacinnabarite was produced - in the laboratory was far from those under which this mineral would be formed in nature. The most successful method is believed to be that by which metacinnabarite was artificially obtained by Allen and Crenshaw (1913). In their experiments the metacinnabarite was produced in relatively macrocrystalline form by the action of sodium thiosulfate on a dilute acid solution of mercury sodium chloride.

But the deficiency of the experiments carried out thus far in the synthesis of metacinnabarite has been that the only criterion for the identification of the products of such synthesis has been the black color, and in a few cases the specific weight, of the products. Only Lehmann (1924), repeating Allen and Crenshaw's experiment, has subjected the metacinnabarite obtained in this manner to X-ray analysis.

To throw some light on the problem of whether metacinnabarite can be of hydrothermal origin, this writer has deemed it useful to carry out experiments in the synthesis of this mineral in the laboratory, under conditions close to those that would have existed in nature in the hydrothermal formation of mercury deposits.

According to the theory of the formation of hydrothermal mercury deposits, the mercury is carried by alkaline sulfide waters in the form of a complex soluble salt of Me_2HgS_2, which is readily formed by the action of concentrated solutions of Na_2S or K_2S on mercury sulfide, according to the reaction:

$$Me_2S + HgS \rightleftarrows Me_2HgS_2$$
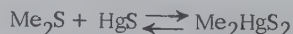
which also results in the formation on the complex anion $[HgS_2]^{2-}$.

As the investigations of Abegg and Jander (1902) and later of Knox (1906) have shown, the solubility of HgS in sulfides of the alkali metals is due to the presence of S^{2-} ions in the solution; these combine with HgS to form the complex ion $[HgS_2]^{2-}$. Since the partial hydrolysis $S^{2-} + H_2O = HS^- + OH^-$ causes HS^- and OH^- to also be present in such a solution, the solubility of HgS in solutions of alkali metal sulfides will depend primarily on the equilibrium between the S^{2-}, HS^- and OH^- ions.

A. A. Saukov, (1946, 1950), in seeking the causes for the precipitation of HgS from alkaline sulfide solutions, based his reasoning on the equilibrium between Na_2S, HgS and Na_2HgS_2, expressed in the following equation:

$$K = \frac{[Na_2Hg\,S_2]}{[Na_2S]\,[HgS]}$$

As the concentration of Na_2S in the solution increases (temperature and pressure remaining constant), the solubility of HgS also increases; on the other hand, when the concentration of Na_2S decreases, HgS will be precipitated from the solution. As factors that would, in nature, cause a decrease in concentration of Na_2S in a hydrothermal solution containing mercury in the form of the complex salt Na_2HgS_2, Saukov cites mainly the following:

1) neutralization of the solution by acids, including CO_2;

2) dilution of the solution by water;

3) the action of oxygen on the solution.

He indicates that all these factors are in a complex interrelationship in nature. When the upward-moving hydrothermal sulfide solutions of mercury encounter downward-moving weakly acidic, neutral or oxygen-rich vadose water, there is a precipitation of mercury sulfide.

In accordance with this scheme, the present writer in his experiments used solutions of Na_2HgS_2, from which mercury sulfides were precipitated by neutralization with acids, or dilution of the solution with water, or by oxidation. The solution of the complex salt Na_2HgS_2 was obtained by dissolving newly precipitated HgS (which occurs in abundance) in a 20 percent solution of Na_2S, after which the undissolved HgS was filtered out of the solution. These freshly obtained solutions of Na_2HgS_2 were used for a number of experiments in the synthesis of metacinnabarite and cinnabar. The products of these syntheses were subjected to X-ray analysis.

Experiment 1.   From a solution of Na_2HgS_2 heated to boiling temperature and partially neutralized by the addition of several drops of dilute H_2SO_4, a black sulfide of mercury (HgS) is precipitated. The precipitation of the mercury sulfide ceased while the solution still contained some quantity of Na_2HgS_2 and gave an

alkaline reaction.   Immediately after the precipitation of the mercury sulfide, the precipitate was filtered out of the solution and washed in the filter. After air-drying, the black precipitate was subjected to X-ray analysis (see table), the results of which showed that the precipitate was metacinnabarite.

Experiment 2. From a solution of Na_2HgS_2, as in Experiment 1, a black mercury sulfide (HgS) was precipitated by partial neutralization with several drops of H_2SO_4.   In this experiment also the solution contained a remnant of Na_2HgS_2 and gave an alkaline reaction.   Thereupon the HgS precipitate, together with the alkaline solution of Na_2HgS_2 remaining above it in the glass, was covered with a watch glass and heated in a sand bath to a temperature of 80°-$90^\circ C$ for 12 hours.

Even after a few hours of roasting it could be observed that the black precipitate of HgS was in places turning to red cinnabar. After further heating, the entire precipitate was changed to the red crystalline modification of HgS -- cinnabar.   This transformation evidently took place because of the different solubilities of these two varieties of HgS in a solution of Na_2S.   The less stable black variety of HgS is more soluble in Na_2S than the red form.   According to Knox's investigations (1906), the solubilities of cinnabar in Na_2S solutions of different concentrations were always less than those of the black forms of HgS.   According to Knox, the ratio of the solubility of black HgS to that of red HgS varied from 1.3 (in 0.1 M solution of Na_2S) to 1.09 M solution of Na_2S).

Thus a saturated solution of the black form of HgS in Na_2S will be supersaturated with regard to the red form of HgS.   The excess HgS is precipitated out as cinnabar, more black HgS goes into solution, and so on.   Hence in the reaction of a saturated solution of HgS in Na_2S on the black variety of HgS, the black form of HgS will gradually be dissolved, and its place will be taken by the precipitation of the less soluble red variety of HgS, cinnabar (Abegg, 1905; Ostwald, 1900).

Experiment 3.   Black mercury sulfide was precipitated out by the addition of a solution of H_2SO_4 to an Na_2HgS_2 solution heated to the boiling point.   In contrast to Experiments 1 and 2, sulfuric acid was added until the alkaline Na_2HgS_2 solution was fully neutralized, and excess H_2SO_4 was added.   Thus all the mercury sulfide was precipitated out, and the solution gave an acid reaction.   The precipitate together with the acid solution in the glass was covered with a watch glass, heated for a week in a sand bath to a temperature of 80° - $90^\circ C$, and then filtered and washed.   X-ray analysis was made of the air-dried black sulfide of mercury. As may be seen from the table, the resulting product turned out to be metacinnabarite.

TABLE 1. Results of

Number of samples	Metacinnabarite (4)		Cinnabar (4)		Product Specimen					
					1		2		3	
					Metacinnabarite		Cinnabar		Metacinnabarite	
	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$
1			4	4.297						
2	10	3.396			10	3.38			6	3.38
3			8	3.346			8	3.346		
4	8	2.943			6	2.93			4	2.92
5			8	2.835			8	2.854		
6			8	2.169						
7	10	2.071	6	2.067	10	2.062	6	2.073	6	2.060
8			6	1.969			6	1.976		
9	10	1.765			10	1.761			6	1.759
10			6	1.745			5	1.737		
11	4	1.690			4	1.683				
12			6	1.672			7	1,677		
13			4	1.579			4	1.578		
14			1	1.495						
15	4	1.455			4	1,457				
16			4	1.432			3	1.427		
17	5	1.343	4	1.339	4	1.338	4	1.339	3	1.338
18	4	1.307	6	1.302	3	1.306	6	1.302	3	1.305
19			6	1.255			5	1.256		
20	9	1.191	4	1,184	8	1.192				
21	9	1.122	5	1,124	8	1.125	4	1.120	6	1.123
22			4	1,102			4	1.103		
23	4	1.030			4	1,032				
24			1	1.074			1	1.069		
25			4	1.033			3	1.030		
26			5	0.991						
27	8	0.998			8	0.988	4	0.998		
28	5	0.274	4	0.976						
29			4	0.057						
30			4	0.940						
31	5	0.926								
32			6	0.919						
33	5	0.887								
34	5	0.842								
35	4	0.816								

NOTES: 1 – The X-ray studies were made by S.S. Levitskaya in the X-ray Laboratory of the 86 mm, v = 30 kV, I = 10 mA.
2 – The interference picture indicates that Specimen 3 is an X-ray amorphous sub-

X-ray analysis

synthesized

				Specimen					
4		5		6		7		8	
Metacinna-barite		Metacinna-barite		cinnabar with metacin, admix.		Cinnabar		Metacinna-barite	
I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$	I	$\dfrac{d\alpha}{n}$
10	3.38	10	3.39					10	3.39
				8	3.346	8	3.350		
6	2.92	7	2.94					7	2.93
				9	2.847	9	2.845		
9	2.062	10	2.069	6	2.067	7	2.065	10	2.065
				7	1.967	6	1.978		
10	1.754	10	1.764	9	1.761	8	1.759	10	1.762
				6	1.737	6	1.732		
4	1.680	3	1.683					4	1.684
				7	1.675	6	1.674		
				4	1.577	3	1.579		
4	1.451	3	1.464					4	1.457
				4	1.431	4	1.429		
5	1.337	4	1.341	4	1.341	3	1.341	6	1.340
5	1.304	4	1.307	5	1.303	5	1.302	4	1.305
				5	1.257	5	1.255		
9	1.191	8	1.194					8	1.191
9	1.124	8	1.124	5	1,126	5	1.126	8	1.124
				4	1,104	4	1.102		
3	1.032	3	1.032					4	1.031
				2	1,067	1	1.067		
				3	1,028	3	1.028		
						5	0.994		
8	0.988	8	0.988					8	0.988
								4	0.974

Department of Crystallography of L'vov University. Data: filtered Fe-radiation, D of chamber=

stance with a small amount of crystalline matter.

Apparently the acid medium favored the preservation of the metacinnabarite for a long period of time.

Experiment 4.   25 ml of saturated Na_2HgS_2 solution at room temperature was diluted in a glass with distilled water to the amount of 500 ml.   Thereupon a black mercury sulfide HgS was precipitated.   The solution above the precipitate gave an alkaline reaction because of the presence of OH^- ions in it.

As mentioned above, only S^{2-} ions are capable of dissolving HgS.   The HS^- ions formed by the hydrolysis $S^{2-} + H_2O = HS^- + OH^-$, unlike the S^{2-} ions, cannot combine with HgS to form a complex soluble salt.   For this reason when the concentration of S^{2-} in the solution is decreased by dilution with water, HgS is precipitated out.

The resulting HgS precipitate was filtered, rinsed, air-dried at room temperature and X-rayed; the table shows that the black HgS precipitate obtained in this experiment was metacinnabarite.

Experiment 5.   25 ml of Na_2HgS_2 solution was diluted with distilled water to the amount of 500 ml.   As in Experiment 4, because of hydrolysis and the decrease in the solubility of HgS, black mercury sulfide was precipitated out.   The solution gave an alkaline reaction. Then the solution plus the HgS precipitated out from it in the glass was covered with a watch glass, placed in a sand bath and heated for eight days to 80° - 90°C. The precipitate was removed from the solution by filtration, and the black HgS thus obtained was washed, air-dried and subjected to X-ray analysis.   The results show that the HgS precipitate obtained in this experiment was metacinnabarite.   Here, even though the solution gave an alkaline reaction, during the course of heating for eight days the metacinnabarite was not even partially changed to cinnabar.   This is explained by the insufficiency in the solution of ions capable of dissolving HgS.

As already noted, the transformation of metacinnabarite into cinnabar is based on the different solubilities of the two varieties of HgS, and thus requires the presence of S^{2-} ions in the solution.

Experiment 6.   A freshly prepared saturated solution of Na_2HgS_2, from which the excess insoluble HgS had been removed by filtration, was allowed to stand in an open glass at room temperature, which did not exceed 23°C. This saturated solution of HgS in Na_2S, containing no excess Na_2S, rapidly dissociated in air with the separation out of HgS. The black HgS precipitate that had separated out and remained under the solution of Na_2HgS_2 at room temperature very quickly, even within a few hours, began to turn into cinnabar.

The table shows the results of X-ray analysis of the precipitate that stood for two days under the solution; this turned out to be a mixture of cinnabar and metacinnabarite.

Experiment 7.   This differed from Experiment 6 in that the black mercury sulfide precipitated out of the saturated Na_2HgS_2 solution was allowed to stand under the Na_2HgS_2 solution at room temperature (no higher than 23°C) for a whole week.   The results of the X-ray analysis show that the resulting product was entirely changed to cinnabar.

The results of Experiments 6 and 7 show that it is not necessary to raise the temperature to obtain cinnabar from solutions, as is often said in the literature.

Experiment 8.   To 20 ml of freshly prepared Na_2HgS_2 solution were added 2 ml of 30 percent perhydrol solution, and the mixture was heated to the boiling point.   The resulting black HgS precipitate was removed from the solution by filtration, washed, air-dried and subjected to X-ray analysis.   The results of the X-ray analysis showed that this product was metacinnabarite.

CONCLUSIONS

In all these experiments metacinnabarite was precipitated directly from an aqueous solution of Na_2HgS_2 by:

1) neutralization of the solution by acid,

2) dilution with water and

3) dissociation of the saturated solution of Na_2HgS_2 in air or by perhydrol.

It is apparent that in nature as well, metacinnabarite is precipitated out first in the hydrothermal formation of mercury deposits.

The subsequent behavior of the metacinnabarite that is precipitated depends ultimately on the medium in which it is formed or with which it later comes into contact.   If the precipitated metacinnabarite is later acted upon by hydrothermal waters carrying alkaline solutions of Me_2HgS_2, it will gradually be transformed into the more stable red form of HgS cinnabar.   But if, after the deposition of the metacinnabarite, the inflow of mercury sulfide solutions ceases or become less, then because of the counteraction of acidic or even neutral vadose waters there will be an insufficient quantity of solvent, in the form of S^{2-} ions necessary to change metacinnabarite to cinnabar and the metacinnabarite may be retained for a great length of time.

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ON THE QUARTZ IN THE BROWN-COAL DEPOSITS
OF DUBROVKA AND THE GLINSK-L'VOV REGION[1]

By

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• *translated by Royer and Roger, Inc.* •

ABSTRACT

The question of new formations of quartz in petrified trees lying in lower Tortonian deposits is dealt with in the article. The mineral substance which replaced the wood is at the present moment all quartz. The microstructural analysis showed that the orientation of quartz was influenced by the walls of tracheids in regard to which the optical axes of quartz are embedded at an angle of 30° - 32°, as well as across the cells of the tracheids. -- auth. English summ.

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The problem of the formation of new quartz in petrified wood has repeatedly been discussed in the literature, (Hoehne, 1954a and b) but has still not been fully clarified. The present writer has made a study of the formation of such new quartz in the deposits of Dubrovka and the Glinsk-L'vov region.

The geologic section through the brown-coal deposits from which the specimens of petrified wood were taken is readily seen in the mine-workings of Dubrovka (around Mt. Rava Russkaya). Here, below the Lithotamnion limestones of the middle stratum of the lower Tortonian stage and upon the eroded surface of Maestrichtian marls, lie the sand-clay-coal deposits of the lower stratum of the lower Tortonian, containing a large number of tree remains in the form of both roots and trunks, apparently carried here by rivers into the littoral zone of the lower Tortonian Sea.

These trees are not all the same: representatives of conifers are more numerous and deciduous trees, less so. Here some trees have been replaced by quartz and are resistant to mechanical influences, while others, unsilicified and turned to coal, readily disintegrate and burn.

Macroscopically, the tree specimens from Dubrovka consist of a dense, highly silicified wood, in places containing remains of the original wood in the form of soft brown fibers. Sometimes such fibers and the walls of cavities are lined with tiny crystals of quartz. In places one may observe fractures crossing the grain of the wood, usually filled with fine-grained dense quartz and druses of tiny quartz crystals. In the preparation of thin sections of such wood,

certain specimens give off an odor of burnt rubber, which has already been mentioned in the literature (Polkunov, 1955).

The specimens from Dubrovka have been analyzed at this writer's request by N. M. Gorshenin.

Under the microscope, in a plane section, tracheids arranged in regular radiating rows are discernible. The annual tree rings are clearly visible because of the sudden alteration of the earlier wood. The sap channels are not observable, but certain of the tracheids have broad bands which in the Miocene evidently acted as horizontal channels for the sap. A large number of ringed pores, forming concentric circles, may be seen in the cell walls. On the basis of these facts N. M. Gorshenin was able to classify these specimens as conifers; the small diameter of the trees suggested to him that they belonged to the species Taxodioxylon taxodii, as noted by the Polish paleobotanist Lilpop (1929). The excellent preservation of the structure of the wood is due to its early mineralization.

The mineral substance that has replaced the wood is at the present time almost wholly quartz; only in some cases does one observe, along the walls of the tracheids, lenticular bands of very small grains of quartz, among which there are also tiny particles of uncrystallized chalcedony with a characteristic zonal extinction.

The diameter of the tracheids in cross section varies from 0.02 - 0.04 mm to 0.1 mm. Sometimes the cell is filled entirely by one grain of quartz, but more often by three or four grains from 0.02 - 0.05 to 0.1 mm in size. The quartz grains often (20 out of 100) have wavy extinction.

This writer was interested in determining what effect the structure of the wood had on the orientation of the quartz. To this end a microstructural analysis was made and the results statistically treated according to the methods of N. A. Yeliseyev (1957).

By means of the Fedorov method (Sobolev, 1954), thin sections were prepared along both the cross sections (figs. 1 and 2) and longitudinal sections (figs. 3 and 4) of the wood and the optical axes of the quartz were determined and plotted on a system of equal-area coordinates. The cross sections were oriented to the annual tree-ring and the medullar ray, whose intersection was the projection of the axis parallel to the axis of the wood; the orienting direction in the longitudinal section was the walls of the tracheid.

FIGURE 3. Longitudinal section through silified wood. One nicol; magnification x 60.
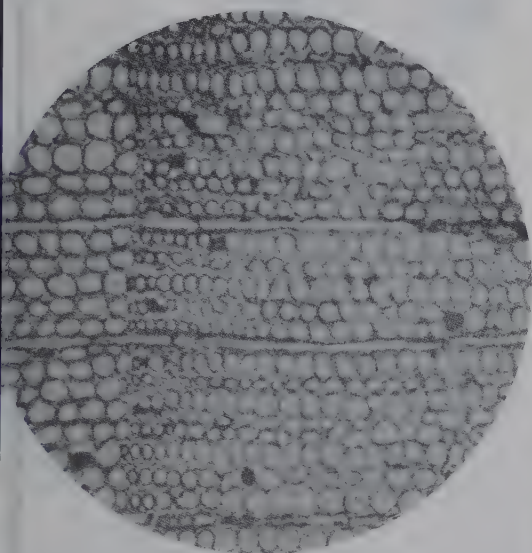
FIGURE 1. Cross section through silicified wood. One nicol; magnification x 60.

FIGURE 4. Longitudinal section through silicified wood. Crossed nicols; magnification x 60.

By drawing isolines for areas of equal point distribution density, orientation diagrams (figs. 5 and 6) were obtained for 500 measurements of quartz grains. These diagrams reveal definite patterns reflecting the orientation of the quartz in the wood.

Regular orientation of minerals in rocks is usually due to deformation. But there are cases, Yeliseyev (1957) has noted, in which such orientation occurs without being associated with any deformation. In such cases one may speak of "tectonite orientation in non-tectonites".
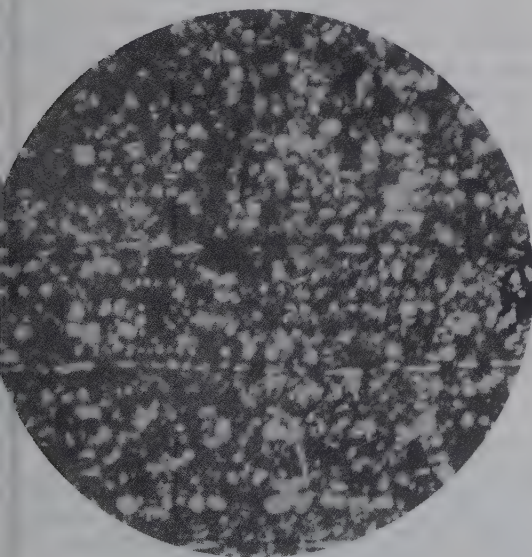

FIGURE 2. Cross section through silicified wood. Nicols crossed; magnification x 60.

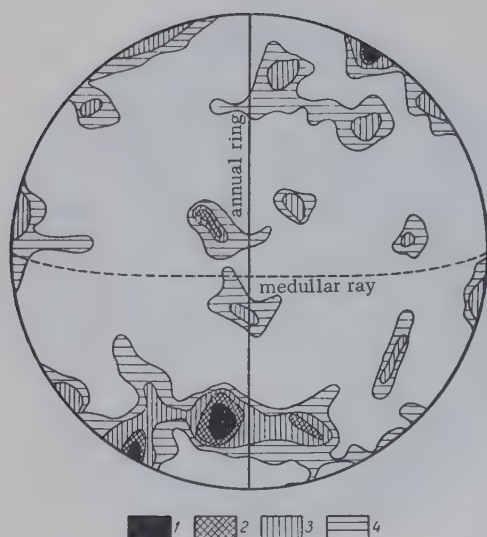FIGURE 5. Orientation diagram of silicified wood for cross section. Density distribution 1 – 56, 2 – 4, 3 – 3, 4 – 2

FIGURE 6. Orientation diagram of silicified wood for longitudinal section. Density distribution: 1 – 8 – 6 – 5, 2 – 4; 3 – 3, 4 – 2

One diagram (fig. 5, cross section) shows two maxima. One, which is fairly intensive, lies in the center of the projection; this indicates orientation of the optical axes of the quartz primarily along the tracheids (along the trunk). The other maximum is marked less intensively; this testifies that the optical axes are also located on the circumference of the projection (transverse to the tracheids) or at a certain angle to it.

The other diagram (fig. 6, longitudinal section) does not contradict the first, although both maxima are more clearly reflected here. The fairly sharp maximum along the trunk indicates that the optical axes are at an angle of 30° - 32° to the walls of the tracheids. The second maximum, which is also quite distinct, reflects a direction of the optical axes primarily normal to the tracheids.

What is the reason for this orientation in the growth of the quartz grains in the wood? The orientation of the quartz has been affected by the walls of the tracheids, which in some cases during the crystallization of the silicic-acid gel have created stresses; as a result of these maximum coefficient of compressibility of the low-temperature quartz (Sobolev, 1957) should coincide with the line forming an angle of about 70° with the optical axis of the quartz — that is, the optical axes of the quartz should lie fairly close to the plane perpendicular to the maximum pressure. But since the stresses in the tracheid cells were evidently not uniform, in other cases the quartz grains grew approximately perpendicular to the walls of the fibers, as in the growth of comb-shaped quartz in veins.

A study of the optical orientation of quartz

grains in petrified wood was undertaken by M. Turnau-Morawska and M. Jahn (1953). They noted that in the petrified wood of Araucarioxylon trees the quartz grains had optical axes at an angle of 30° to the tracheid walls. To establish the optical orientation of the quartz grains these authors used the method of finding the angle of extinction between the direction of the trunk's axis and ϵ for 200 grains in the longitudinal section (without the Fedorov stage).

The present writer's data, obtained by measuring quartz grains on the Fedorov universal stage in thin sections prepared both in the longitudinal and the cross sections, confirm the results of the study just mentioned above. It was also established that the optical axes of the quartz in the petrified wood from Dubrovka are at an angle of 30° - 32° to the walls of the tracheids (the axis of the tree trunk).

Since they did not make measurements on the Fedorov stage, the above authors were unable to note the second maximum (transverse to the tracheids), which in the diagrams cited in this article is just as clearly developed as the first, indicating an orientation along the tracheid cells.

The specimens of wood from Glinsk (around Nesterov) are poorly preserved, but in thin sections in the cross section one may discern microscopic features typical of larch trees with ring-like pores. Here the nature of the quartz formation differs somewhat from that in the trees from Dubrovka, where the quartz grains are usually irregular or repeat the shapes of the wood cells.

The specimens from Glinsk also show inten-

sive quartz formation, but here the annual rings have been noticeably exfoliated and the wood is solidly packed with a multitude of tiny quartz crystals (fig. 7). It can be seen macroscopically that the surface of the exfoliated annual rings is most densely filled with double-terminated quartz crystals, producing a strong glassy lustre. These tiny crystals are generally randomly oriented, but frequently lie with their long side parallel to the plane of the annual rings, and sometimes at an angle to it, so that one termination of the quartz crystal appears to be em-

FIGURE 7. Quartz crystals in silicified wood. Magnification x 10.
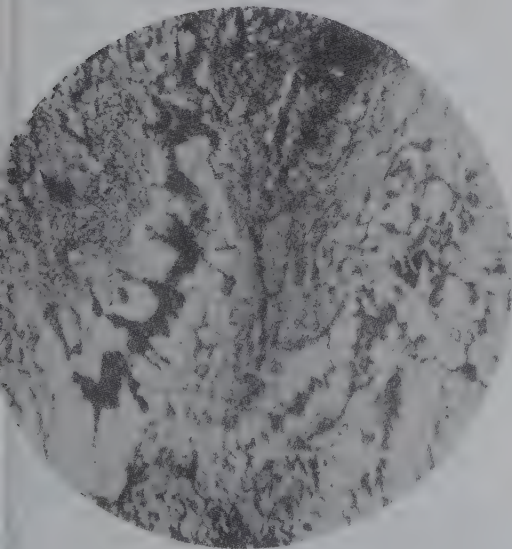
FIGURE 8. Crustification structure of quartz. One nicol; magnification x 25.
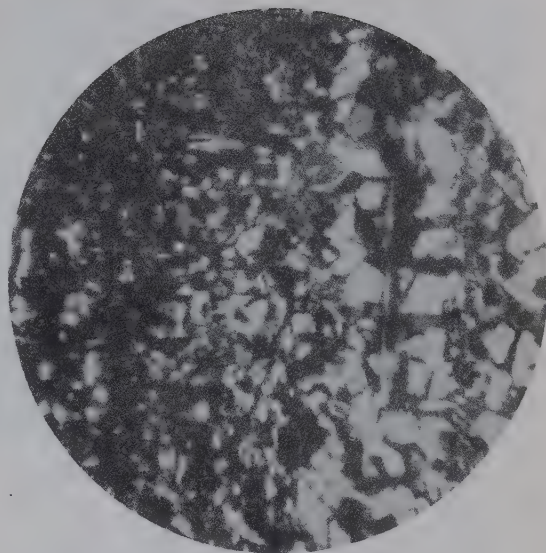
FIGURE 9. The same as Figure 8, with crossed nicols.

bedded in the wood. Under the microscope in the longitudinal sections one may see wood fibers and a disorderly aggregate of elongated quartz grains from 0.02 to 0.074 mm in length. The quartz cystals show no wave-like extinction.

In some places one may observe crustification structures (figs. 8 and 9); here the fiber has formed a wall for the growing crystals, and one can see that the crystals growing opposite to each other (perpendicular to the fiber) have pushed the fiber apart.

This writer has studied quartz crystals up to 1 mm in size. These are combinations of a hexagonal prism and two rhombohedra. The rhombohedra — right and left — have almost identical developments; sometimes they are full of gaps. The crystal faces have transverse striations; gaseous-liquid and liquid inclusions were not observed.

The quartz crystals are colorless and transparent, but the rhombohedra are colored brown; it may be supposed that in their growth the tiny crystals encompassed a certain amount of organic matter that now forms hour-glass shapes in them (Lemmleyn, 1951). (fig. 10).

A heating curve (fig. 11) was obtained for these crystals by heating them in an electric furnace and weighing them at 100-degree intervals. The heating curve shows that at the temperature of 450° the organic substance has been totally burned up. At 400° the organic matter within the crystal seems to become dark and to diffuse, so that almost the whole crystal becomes black (only at the very center is there a small remnant of transparent area). At 450°

FIGURE 10. Tiny quartz crystals from Glinsk.
Magnification x 15.
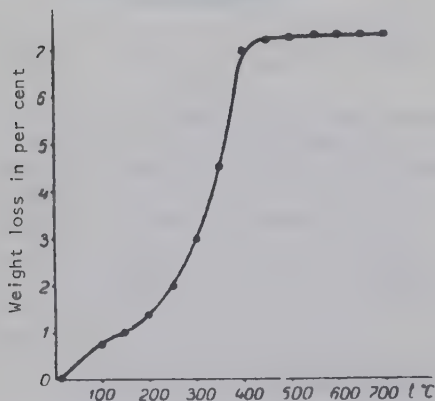
FIGURE 11. Heating curve of quartz crystals.

FIGURE 12. Quartz crystals after roasting at
450°. Magnification x 15.

the crystals become colorless and transparent; the rhombohedra on the other hand have become almost completely gray and opaque (fig. 12).

Study of the wood specimens altered to quartz suggests that the process of quartz formation did not take place all at one time, but in various periods in the formation of the brown-coal deposits.

Silicification of part of the Dubrovka wood was completed before the process of coal formation, so this wood is completely preserved and the mineralization took the course of metasomatic replacement of the wood. Probably the initial solutions penetrating the wood were colloidal, as indicated by the rare grains of non-crystalline chalcedony, as well as by the fractures that run normal to the grain of the wood and may owe their formation to the very great stresses associated with recrystallization of the siliceous gel.

The formation of the quartz crystals in the Glinsk wood occurred later, in a soft and friable woody mass, after the structure of the tree had disappeared. This was followed by stable conditions that governed the formation of the quartz crystals and the development of rhombohedral terminations at both ends. These crystals, most quickly, were formed from true solutions. The tiny transparent crystals of quartz in the cavities and fractures in the wood from Dubrovka probably also grew from true solutions.

The question inevitably arises: what was the role of the wood in the formation of the quartz in the brown-coal deposits? The trees were evidently good channels and reservoirs for the mineral solutions and facilitated their concentration, as well as having an orienting effect on the growth of the crystals. On the other hand, as I. Ye. Leskevich (1952) points out, the quartz crystals may have arisen in the brown-coal stratum as well, forming lenses of microcrystalline quartz there.

It must be noted once more that for some reason in one and the same deposit, and even in the same mine shaft, one encounters both silicified and nonmineralized, highly carbonized trees. The reason for this is still hard to explain: it can only be assumed that the mineralization did not occur everywhere and that the solutions apparently traveled along particular directions. The occurrence of silicified and carbonized trees in the deposit should be examined from this standpoint.

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SOME LAWS GOVERNING THE REGIONAL GEOLOGIC OCCURRENCE OF FERROUS AND NONFERROUS METAL ORES[1]

by

Ye. Ye. Zakharov[2]

•   translated by Paul T. Broneer, Royer and Roger, Inc.   •

ABSTRACT

This article surveys the geographical distribution of metal ores, grouped into six categories: iron, manganese, chromium, silver-lead-zinc, copper and tin.   The author reviews the occurrence of these ores throughout the world, and their association primarily with broad geotectonic zones. The essential information is encompassed in six maps.   Legends explain the classification of the deposits by age (metallogenic eras), type and paragenetic associations of minerals.   The symbols for these various classifications of ores are superimposed on maps showing the zones of Precambrian (cratonal), Paleozoic, Mesozoic and Cenozoic folding.   The particular interest of this paper is that the author has made this survey as a preliminary to dividing the earth's surface into metallogenic provinces (of which he distinguishes 21), giving a brief discussion of each and showing its connection to areas of different ages and geotectonic structures.   The ultimate aim is to construct "prognostic maps", or maps drawn on such a basis that they may be used to predict the occurrence of a given ore or ores in a given region. --Paul T. Broneer.

* * *

One of the most important problems in the study of ore deposits is the discovery of the general regional geologic laws governing the world distribution both of the individual types of ore deposits of a given metal and of the combinations of genetic types of ore deposits of all metals that are known at the present time.   The discovery of these laws and their charting on diagrams will make it possible to construct maps of the metallogenic provinces of the world and to solve many problems in the theory of ore formation.

During the last quarter century many papers have been published, both in the Soviet Union and other countries, in which attempts have been made to determine these laws for the territories of individual states, regions and districts.   Geographic maps of the world showing the distributions of the largest ore deposits have been included in surveys of the individual types of raw materials and in certain scientific texts.   Moreover in recent years the staffs of VSEGEI, IGEM of the Academy of Sciences of the U.S.S.R., GIN of the Academy of Sciences of the Kazakh S.S.R. and other scientific organizations have done much work in preparing maps forecasting the occurrence of metal ores in individual districts of the Soviet Union.

At the present time the preparation of such forecasting maps has already achieved some definite positive results both in elaborating the

method of preparing such maps with their symbols, and in their interpretation.   This cannot, however, be said of the metallogenic maps of the world.   In the case of the latter, it is evidently necessary to work out a special method of preparation, special systems of designations and symbols, and so forth.   In this it is scarcely possible to use the experience in preparing forecasting maps (which are mainly large-scale - from 1:50,000 to 1:500,000, and more rarely 1:1,000,000) in elaborating methods of preparation and systems of symbols and designations for metallogenic maps of the world (which are mainly on a small scale - from 1:5,000,000 to 1:25,000,000) On such small-scale maps it is technically impossible to reflect the connections between the individual ore-bearing zones and other forms of occurrence of mineralization with specific elements of geologic structure, stratigraphic and lithologic strata, intrusive massifs, etc. A system of designations must be worked out in which the respective symbols will indicate not only the ore deposit or group of deposits of a definite mineral association, but also their structural geologic position, their occurrence with a particular intrusive or in the rock overlying it, the nature of the alterations in the host rocks, etc.   In other words, the given symbol must reflect a definite ore formation (or genetic type of ore deposits) which will be a deposit of a single type characterized by specific complexes of genetic features present only in this formation, and particularly by the specific geologic conditions of the occurrence and morphology of the ore bodies, which are very close to the paragenetic mineralogical associations and to the rocks enclosing the ores, etc., showing that the formation and occurrence of such an ore body in the particular part of the earth's crust is determined by a favorable combination of geotectonic, igneous, facies and lithological, geochemical, physicochemical and physical geographic factors (Zakharov, 1953, 1955).

Another complicated problem in preparing metallogenic maps of the world is the selection of a base map on which the respective symbols designating the ore formations will be placed. As one possible variation of base maps, the present writer has chosen an extremely simplified diagram of the geotectonic regionalization of the world. In the future, however, it will obviously be necessary to develop a special cartographic base map for these purposes. Maps showing the distribution of ore deposits were prepared for the individual and most widespread ferrous and nonferrous metals. It is assumed that the empirically discovered laws governing the emplacement of each metal will ultimately make it possible to prepare a combined metallogenic map of the world.

To this article have been added maps[3] (figs. 2-6) showing the distribution of ore formations of iron, manganese, chromium, silver-lead-zinc, copper and tin, in the preparation of which the present author has taken account of the above-mentioned circumstances. Proceeding from his assumption that the geological or geotectonic factor is one of the most important in ore formation, the author, as already indicated, has taken a simplified geotectonic map as his base. In the specially developed system of designations, the symbol for the specific forms of occurrence correspond (as much as possible) to the respective genetic groups of deposits of all the metals.

The most important regional geologic laws governing the distributions of these ore deposits are set forth below.

REGIONAL GEOLOGIC DISTRIBUTION OF IRON ORE DEPOSITS

The distribution in space of ore deposits of deep-seated origin is determined by a combination mainly of geotectonic, igneous, geochemical and physicochemical factors, whereas in the distribution of ore deposits of surficial origin the most important roles are played by combinations of geotectonic, facies and lithologic, geomorphologic, geochemical, physicochemical and geographic (climatic) factors. For this reason one must examine separately the laws governing the distribution of iron ores of deep seated and surface origin.

Hypogene iron ore deposits are associated with geosynclinal regions. An especially large number of them are concentrated in areas in which the oldest folding has occurred, and recent igneous activity is manifested only in connection with deep faults in the earth's crust.

The specific character of these igneous intrusions to a considerable degree predetermines the genetic types of iron-ore deposits associated with them. These are gabbro-norite and anorthosite intrusive massifs, which contain deposits of titanomagnetites and titaniferous magnetite ores. The map (fig. 1) shows that the overwhelming majority, and especially the largest properly igneous titanomagnetite deposits, within the massifs of basic intrusive rock are associated with ancient platforms - shields. For example, the deposits of Mine Center, Ollard Lake, Ivry and Saint Urbain (Quebec, Canada), Iron Mountain and Sheep (Wyoming, USA) and others are located in the Canadian shield. The deposits of Rodsand and Soggendal (Norway), of Taberg and Rutivar (Sweden) and of Vuolijoki (Finland) are located on the Scandinavian shield. The numerous titanomagnetite deposits of Tanganyika, the Union of South Africa, Togo, Ghana, Kenya, Sierra Leone and other states are located on the African shield. Moreover similar deposits are known on the Brazilian and Indian shields.

Within the Siberian platform, however, titanomagnetite deposits of this type have yet to be found. Here there is a widespread distribution of magnetite and magnetite-hematite ores in the form of veins and stocks which are evidently postmagmatic deposits genetically associated with gabbro diabases (traps). The latter are apparently the same as the gabbro-norite intrusives of other shields. In many cases the traps have been crystallized under hypabyssal and sometimes even extrusive conditions, which may have determined the genetic features of the iron-ore deposits of the Siberian platform.

Among other genetic types of iron-ore deposits of deep-seated origin observed on ancient platforms, mention must be made of the apatite-magnetite deposits in syenites. A. N. Zavaritskiy has suggested that they have a complex synthetic-liquation origin. This interpretation of their genesis corresponds to the somewhat higher alkalinity of the extrusive rocks containing the ores. There is also a possible close genetic or paragenetic association between these rocks and the complexes of alkaline rocks that are frequently observed under the same geotectonic conditions, often in local association with massifs of gabbro-norite rocks (at Khidiny and Monchetundra on the Scandinavian shield, Bushveld on the African shield, etc.).

Deposits of magnetite formations in skarns are encountered much more rarely on ancient platforms. Reliable data on the age of all these gabbroidal, syenitic and alkaline intrusives on shields are lacking. It may be supposed that they are partially Caledonian, and perhaps Hercynian. No doubt some of them are still older - Precambrian -- and the iron-ore deposits genetically associated with them are of equal age.

The original maps in the book are in color, at a scale of approximately 1:50,000,000. They have been modified for black and white presentation and reduced in this translation.--M. R.

For delineation of Map Units I-V, see Figure 3

☐ I ▨ II ☐ III ☐ IV ☐ V

Ores of igneous origin proper:

✚ Titanomagnetites in gabbro
✚ Apatite-magnetite and magnetite in syenites and granites

Post-magmatic deposits:

▲ Magnetite in skarns
△ Magnetite-hematite
● Metasomatic siderite

Deposits resulting from weatherings:

◆ Nontronite-goethite (weathered crust of ultrabasic rocks)
◆ Opal-hydrohematite hydrogoethite (oxidized gossan of pyrite deposits)
◇ Hydrogoethite-goethite (infiltration in karsts)

FIGURE 1.   Distribution of

This is confirmed by the considerable dynamic metamorphism undergone by certain of the titano-magnetite deposits in gabbro and magnetite skarns (in Africa and the Scandinavian peninsula). At the same time, there are fairly convincing geologic data on the younger age of certain of these intrusive igneous formations on platforms. For example, the gabbro-diabase (traps) have altered the late Paleozoic and even Jurassic deposits that compose the upper part of the sedimentary mantle of the Siberian platform.

The nature of the hypogene iron-ore minerali-zation in regions in which Alpine, Hercynian and Caledonian folding is developed differs some-what from the corresponding mineralization on ancient platforms.   It has already been mentioned that a comparatively large number of hypogene iron ore deposits are associated with regions in which Alpine geosynclines and zones of Alpine folding are developed.   As one may see from the map (fig. 1) hypogene iron ore deposits associated with these structural zones form
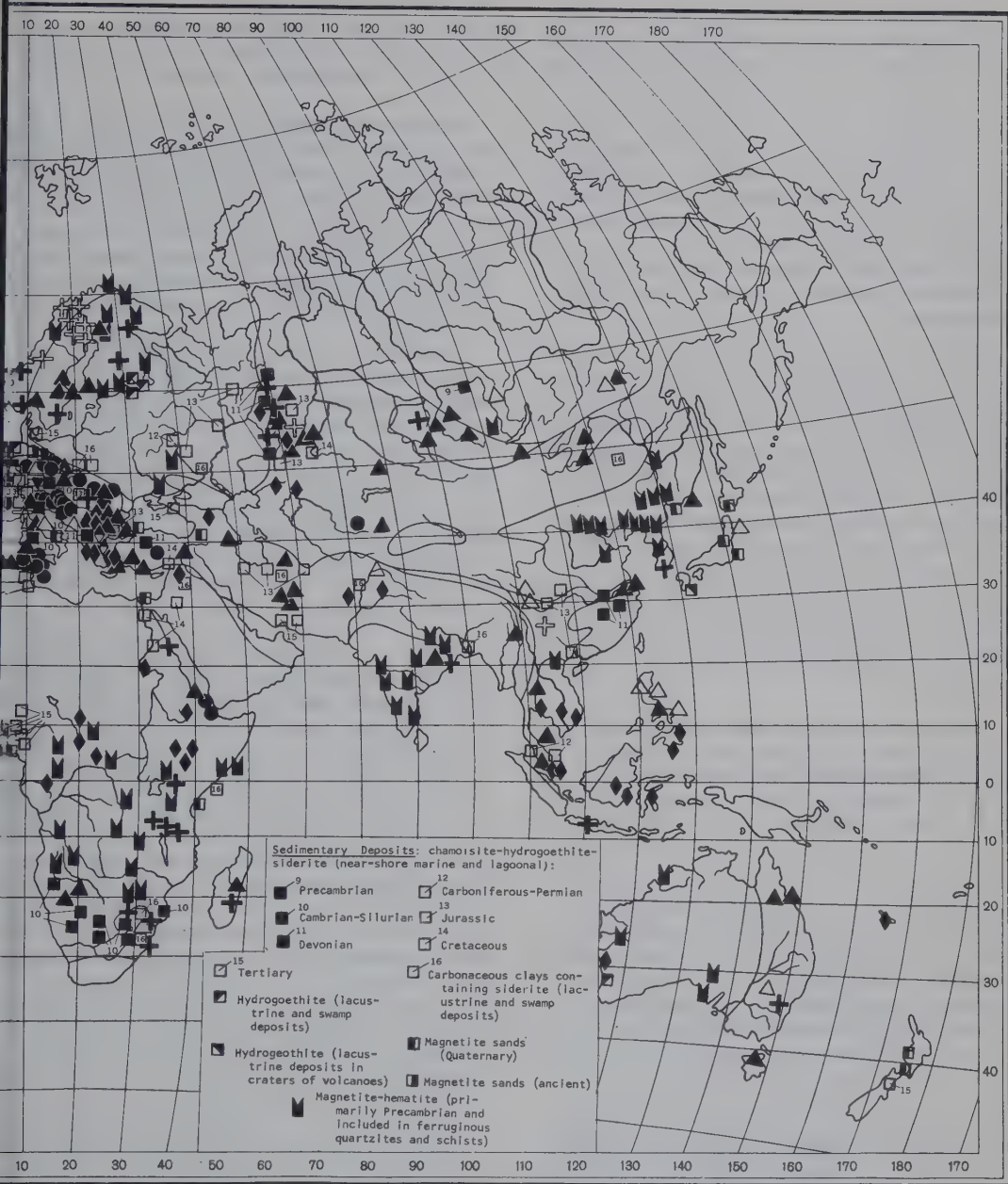
the world's iron-ore deposits

Legend within the map:

Sedimentary Deposits: chamoisite-hydrogoethite-siderite (near-shore marine and lagoonal):

9 Precambrian
10 Cambrian-Silurian
11 Devonian
12 Carboniferous-Permian
13 Jurassic
14 Cretaceous
15 Tertiary
16 Carbonaceous clays containing siderite (lacustrine and swamp deposits)

Hydrogoethite (lacustrine and swamp deposits)

Hydrogeothite (lacustrine deposits in craters of volcanoes)

Magnetite sands (Quaternary)

Magnetite sands (ancient)

Magnetite-hematite (primarily Precambrian and included in ferruginous quartzites and schists)

---

elongated belts along the shores of the Pacific Ocean on the continents of America and Asia, and also form a zone extending from west to east, almost parallel to the equator, from the Iberian Peninsula and the extreme northwest of Africa through southern Europe, Asia Minor and Central Asia to China and India. These deposits are most frequently represented by magnetite formations in skarns, magnetite-hematite formations and siderite metasomatic formations. Magnetite deposits of igneous origin proper, genetically associated with syenites and granite, are encountered much less frequently.

Within the Alpine folded zones one may note certain regularities in the distribution of various hypogene types of iron-ore deposits. For example, the zones of Alpine folding in Europe and North Africa are characterized by a very extensive development of deposits containing metasomatic siderite ores, whereas deposits

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of this genetic type are almost completely lacking in the zones of Alpine folding that surround the Pacific Ocean. On the other hand, in the latter there is a great development of magnetite in the skarns. In general, skarn iron-ore deposits of Alpine age are most frequently genetically associated with quartz-diorites and granodiorites. Apparently many of these magnetite ores in skarns were formed at comparatively small depths. This is indicated by the direct transition from extrusive rocks into hypabyssal facies of effusive rocks that have taken part in the processes of skarn formation observed in the deposits.

The iron-ore deposits of deep-seated origin that are of Hercynian age are generally represented by the same genetic types of those of Alpine age, but in Central Europe the predominant development is that of siderite metasomatic deposits. Central and Southern Europe and North Africa together thus form a peculiar iron-ore metallogenic province of deep-seated mineralization, in which two metallogenic epochs have been noted: Hercynian and Alpine. The predominant genetic type formed in both these epochs is the metasomatic siderite formation. The reason for the widespread development of deposits of the siderite formation is still not clear.

There are important differences in the nature of the hypabyssal iron-ore mineralization in the zone of Hercynian folding in the Urals. This is a peculiar geologic province in which, in contrast to many other regions in the Paleozoic, there were favorable conditions for differentiation in the magma chamber which determined the formation of the large gabbroic massifs. The latter contain numerous titanomagnetite deposits and differ somewhat from the ore-bearing massifs of gabbroic rocks developed on ancient platforms, where they are represented by gabbro-norites, hyperites, etc. Apparently the nature of the Ural gabbroic massifs and of the titanomagnetite mineralization genetically associated with them was determined by the conditions of their formation in the zone of Hercynian folding that developed in the corresponding geosynclinal region. At the same time, one must note certain similarities between them. In particular, in the Urals as on the Scandinavian shield, there are known to be deposits of apatite-magnetite and magnetite formations (although these are very small).

These deposits are igneous formations genetically associated with syenites, and the latter in turn are associated by differentiation with the gabbros.

One of the most widely developed genetic types of iron-ore deposits in areas of both Hercynian and Caledonian folding is magnetite in skarns genetically associated with syenites and granites. Hydrothermal deposits of magne-

tite-hematite and metasomatic siderite are much less frequently found. As already noted, one exception to the latter generalization is the region of Hercynian folding in Central Europe.

Now let us consider the distribution of hypergene iron ore deposits. Among these, the most widespread and largest concentrations of metal are those of sedimentary origin and deposits associated with the formation of a weathered crust on ultrabasic and basic rocks. For this reason the laws governing the emplacement of hypergene iron ore deposits of the two main types indicated above will be considered.

According to N. M. Strakhov (1947) in studying regional tectonic localization of sedimentary iron ore minerals, it is convenient to distinguish three categories of structural members in the earth's crust: platforms, geosynclines and regions of previous (relative to a given moment) folding, such as Caledonian structures in the Hercynian tectonic age, Hercynian structures in the Alpine age, Laramide structures in the Cenozoic, Alpine structures at the end of the Neogene and in Quaternary times.

Strakhov, on the basis of much data, makes a number of generalizations on the distribution of sedimentary iron ores. He notes that iron may be accumulated under different tectonic conditions in geosynclines, in regions of previous folding and on platforms. Nevertheless, masses of iron ores differ sharply within different tectonic structures; for example, sedimentary near-shore marine iron-ore deposits are formed mainly in geosynclines, although deposits of this genetic type are encountered in all types of structures; lacustrine and bog ores, which are also developed under all structural conditions, are clearly concentrated on platforms and in regions of past folding; iron-ore deposits originating in a weathered crust are connected primarily with regions of previous folding and with platforms. It is therefore suggested that they are absent in geosynclinal zones (in the stage of subsidence).

Thus the principal regional tectonic elements of the earth's crust are characterized by an extremely variegated development of different facies and types of sedimentary iron ores within them. Moreover, Strakhov notes that in the development of any type of iron ore in various tectonic parts of the earth's crust, its chief mineralogical and petrographic features are preserved without change; this may be illustrated by many examples. On the other hand, it cannot be said that ore formations of a single type are identical in different tectonic zones.

Strakhov attaches great importance to the geochemical properties of these types of ores, and in particular to the differences in the content of such trace elements as copper, chromium

alt, nickel, arsenic, and vanadium. On
l basis of the existing analytical data (al-
ugh these are still not very numerous) he
gests that iron ores originating in regions
re the folding has recently come to an end
characterized by the greatest variety and
greatest concentration of trace elements.
aller variety and concentrations of chemical
mixtures typify geosynclinal ores.

in regard to their content of admixtures of
ce elements, the ores of platform regions
upy an intermediate position between ores
olded zones and those of geosynclines.
akhov explains this by the fact that in zones
previous folding, during the process of fold-
and for a long time afterwards, erosion
uncovered and carried away the upper ore-
ring parts of the intrusion and their contact-
tamorphic aureole. The above-mentioned
mical elements in these rocks have gone into
ution (although, evidently, only in part),
re migrated in surface waters and finally
led in the hypergene iron ores being formed
he area of folded zones. Platform ores are
rer in admixtures, because such regions
a rule contain the deeply eroded roots of ore-
ring intrusives.

in Strakhov's opinion, the geotectonic locali-
ion of the iron ore deposits has been de-
mined mainly by the following:

a) the paleogeographic conditions charac-
izing platforms, geosynclines and regions
previous folding;

b) the regime of dynamic movements pres-
in each of these structural areas; and

c) the lithologic composition of the substrate
ming the surficial part of the lithosphere.

Until recently there were no broad and
austive investigations, on the basis of
ich one might characterize platforms, geo-
clines and regions of previous folding from
the above points of view. The first publica-
n of this nature was N. M. Strakhov's mono-
aph, "Iron Ore Facies and their Analogue in
History of the Earth" (1947).

Before considering such of the world's
n-ore metallogenic provinces (hypergene
neralization) as are now known, we must
cover how much continuity there is in the
dencies of the formation of hypergene iron-
mineralization throughout geologic time.
om the Cambrian to the Quaternary alone,
re are no less than six major and eight or
e minor iron-ore metallogenic epochs.
ese have alternated with periods in which no
n-ore deposits, for all practical purposes,
re formed. No doubt future studies will
plement this information with new data on
metallogeny of hypergene iron-ore deposits.

Strakhov (1947) distinguishes the following
metallogenic epochs in sedimentary iron-ore
mineralization: 1) Lower Cambrian;
2) Arenigian-Llandeilian; 3) Llandoverian;
4) Lower Devonian; 5) Middle-Upper Devonian;
6) Lower Carboniferous; 7) Upper Carbonifer-
ous; 8) Permian; 9) Jurassic (mainly Liassic-
Dogger); 10) Neocomian; 11) Emscherian;
12) Cenomanian; 13) Danian-Eocene; 14) Neo-
gene (Pliocene).

Among these epochs the most important are
the Jurassic, Neogene, Upper Carboniferous
and Cambrian-Silurian. It must be kept in mind
that during each of them, including those with
very intensive mineralization, iron-ore was
deposited in only a relatively small number of
regions of limited extent, the iron-ore metal-
logenic provinces. It must also be kept in
mind that the entire territory of a metallogenic
province is not completely filled with sedimen-
tary ore, but is composed of certain areas of
relatively large concentrated deposits. At the
same time, outside the boundaries of a province
one sometimes encounters isolated deposits
which represent either the members of a new
and still incompletely discovered iron-ore
province or else the vestiges of a province
which has to a considerable degree been re-
moved by erosion.

Let us now examine the principal metallo-
genic provinces of hypergene (sedimentary)
iron-ore mineralization of the world. Hyper-
gene iron-ore deposits of the Cambrian and
Silurian are localized in certain regions
(see fig. 1).

The North American metallogenic province
is represented by concentrations of iron ores
in Virginia, Pennsylvania, Georgia, and Ala-
bama; here the ore is contained in a thick
quartz series of the Lower Cambrian and is
represented by oölitic hematite-chamoisite-
siderite ores; and in Newfoundland and Nova
Scotia by oölitic marine hematite-chamoisite-
siderite of Arenigian-Llandeilian age.

The Western European metallogenic province
is represented by iron-ore deposits in the
northeastern part of Portugal, in Ordovician
quartzites and shales (the deposits of Guadra-
minal and Monkornu); by sedimentary iron-ore
deposits in northwestern Spain (Asturias, Leon
and Galicia) in Cambrian and Silurian rocks
and characterized by hematite-leptochlorite
ores; by sedimentary iron-ore deposits of
Normandy, Brittany and Anjou in France,
which are oölitic chamoisite-hematite ores in
Lower Silurian rocks; by deposits of Czecho-
slovakia (Nučić) where the iron ores occur in
quartzites and graywackes of Arenigian-
Llandeilian age; and by sedimentary iron-ore
deposits of the same age in Italy (on the
island of Sardinia and in the Alps).

The South African metallogenic province includes a large number of iron-ore deposits in the Union of South Africa (in the districts of Hemagar, Pretoria, and the Transvaal), Swaziland and Southwest Africa. These deposits are represented mainly by oölitic hematite ores of Cambrian and to a lesser degree of Silurian (Karroo system) age. The relative age of these deposits is a matter of dispute; certain geologists assign the greater part of these deposits to the Precambrian.

In the South American metallogenic province Argentina contains a number of sedimentary iron-ore deposits (Sapla, Norta, Sur), in sandstone deposits of Ordovician age. This suggests there is a metallogenic province in this region. Hypergene iron-ore deposits of the Devonian period are located in a number of provinces.

The East European metallogenic province is represented by numerous sedimentary marine iron-ore hematite-chamoisite deposits ranging along the western slopes of the middle and southern Urals; this province stretches westward into the Russian platform (in the form of interbeds of oölitic iron ores in Devonian deposits of the Tumaza and Novokhopersk regions).

The West European metallogenic province is represented by sedimentary hydrogoethite deposits of Devonian age, in the districts of Campin, Namur, Couvin and Liege in Belgium. Certain geologists attribute an extrusive-sedimentary origin to a number of iron-ore deposits in Germany, Czechoslovakia and other countries. Here mineralization is associated primarily with the boundary between the Middle and Upper Devonian. If the latter deposits actually have such an origin, the metallogenic province must also extend over those territories.

The Chinese metallogenic province is represented by oölitic hematite ores occurring in sandstone and shale series of the Devonian in Hunan, Tsiangsi [Shansi?] and Hopeh provinces.

Hypergene iron ore deposits of Carboniferous age are located mainly in Europe, where one may distinguish two metallogenic provinces.

The East European province is represented by a comparatively small number of deposits of Lower Carboniferous age; these include the Tula iron ore deposits, the siderite coal-bearing series of the Izelov, Kynov and Serginsk districts in the Urals and certain other places.

The West European Carboniferous metallogenic province has iron-ore deposits in coal-bearing series of paralic basins in England and nearby regions. These are mainly siderite and hydrogoethite deposits of Upper Carboniferous age in Ireland, Scotland, Wales, Nottinghamshire, Derbyshire and other areas of England.

It has already been mentioned that the Jurassic included one of the principal iron-ore metallogenic epochs; in this epoch were formed the largest iron-ore deposits of the world. Hypergene iron-ore deposits of Jurassic age are located in at least four metallogenic provinces.

The West European province is the largest; here during the Liassic, Dogger and part of the Malm were formed enormous masses of marine oölitic "minetta" ores, whose composition very closely resembles that of the oölitic hematite-chamoisite-siderite ores of the Cambrian and Silurian periods. These deposits are spread over great areas of France, Luxemburg, Germany and Belgium; this group must also be taken to include the deposits of Cleveland and other similar deposits of England and the sphene-siderite Jurassic deposits of Poland.

The Northern Eurasian province is the second metallogenic iron-ore province of the Jurassic; this is located entirely within the Soviet Union, in both its European and Asiatic parts. This province is represented by the hydrogoethite of the Litsepskiy iron-ore basin, the siderite-hydrogoethite deposit of the Gorkiy and Kirov districts, the siderite-hydrogoethite deposit of the eastern slopes of the Middle Urals and the hydrogoethite-leptochlorite-siderite deposits of the southern Urals and Kazakhstan; it may also include certain deposits of western Siberia.

The Balkan-Asia Minor province has been called the Mediterranean province by N. M. Strakhov; he indicates that geographically this province coincides with the zones of distribution of the Lower and Middle Jurassic deposits in the marginal areas of the Mediterranean orogenesis. This province includes deposits of oölitic and rich silica-hematite ores among the Middle and Upper Liassic sediments in the Bukhovo-Lokar and the Gradetskiy-Balshan regions in Bulgaria, and also a number of deposits of argillaceous iron ore among the Jurassic coal-bearing sediments of Iran.

The Southern Chinese province in general coincides with the distribution of the coal-bearing series at the end of the Triassic and in the Jurassic period. Iron-ore deposits of Jurassic age represented by marine oölitic hematite ores occurring at the base of the Tszymotszynskaya suite are known only in Southern Szechuan, but throughout Szechuan and many other provinces of China the Jurassic coal-bearing strata include many bog-iron ores.

In contrast to the Jurassic, the Cretaceous is characterized by somewhat less intensive accumulation of iron ore, but concentrations of iron-ore deposits were formed throughout large areas.

Three metallogenic provinces in which there ere comparatively more intensive processes of ypergene iron-ore mineralization during the retaceous are mentioned below.

The North African metallogenic province is epresented by sedimentary iron-ore deposits French Morocco; by deposits of oölitic iron res of Cretaceous age at Dzhabel-Ank, Dzhabel-nombi and other places in Tunisia, and by litic iron ores in ferruginous sandstones of pper Cretaceous age (Cenomanian) in Egypt. his metallogenic province must, apparently, clude the sedimentary iron ores among the retaceous sediments in Transjordan and the litic Cretaceous iron ores on the southern past of Anatolia in Turkey.

The Western European metallogenic province represented by sedimentary lacustrine iron-re deposits of Cretaceous age at Amberg, ullbach and Auerbach in Germany, the iron res in the clastic Cretaceous deposits in the altzgitter and of Gross-ilsed-pein in Germany; nd the Lower Cretaceous iron ores of England nd elsewhere.

The Northern Eurasian province is similar the Jurassic metallogenic province and located ntirely within the Soviet Union. It is repre-ented by the Khoper Cretaceous iron-ore de-psits in the Voronezh Oblast; by similar iron re deposits of hydrogoethite-chamoisite-sider-e composition on the eastern slopes of the rals and in the Kustanay Oblast in Kazakhstan.

As mentioned before, a concentration of iron ineralization again took place in the Cenozoic. ypergene iron ore deposits of the Tertiary re grouped in at least five metallogenic prov-ces.

The West European province is represented y oölitic iron ores in Tertiary deposits at iddelfart in Denmark; by iron minerals in the ower Pliocene ferruginous sandstones at Distien Belgium, by Eocene iron-ore deposits at ressenberg and Gruenten in Germany, and by ocene pisolitic deposits in Switzerland and outhern Germany.

The Crimean-Caucasus metallogenic prov-nce is represented by middle Pliocene hydro-oethite-chamoisite-siderite ores of the Kerch ron-ore basin, encompassing the Kerch and aman peninsulas.

The Southern Iran province is represented by edimentary iron-ore deposits of Tertiary age oncentrated along the shores of the Persian and man gulfs and on the islands of Khormuz and arur.

The American province is represented by arine hydrogoethite-chamoisite ores of middle ocene age in Texas and Alabama. The com-mercially significant deposits are concentrated mainly in the coastal plain of Texas.

The West African (Nigerian) metallogenic province is represented by numerous sedimen-tary iron-ore deposits of Upper Cretaceous-Paleogene age in Nigeria and directly adjoining regions. The deposits are characterized by marine oölitic hydrogoethite ores.

As noted before, iron-ore deposits originating in the weathered crust are associated primarily with regions of previous folding and platforms. According to N. M. Strakhov, iron ores of this genetic type, like laterites in general, are con-nected with a more-or-less clearly formed peneplain. The majority of investigators re-late the formation of these ores to the end of a lengthy continental period before the beginning of a new erosional cycle, a time in which pene-planation or conditions very close to it pre-dominated. For example, A. D. Arkhangelsky (1937) dates the weathering crust of the Urals, Kazakhstan and other adjoining regions as early as Mesozoic. Strakhov considers that the weathering ores were formed exclusively in a topography that was gradually eroded and finally acquired a completely peneplained sur-face; such conditions are extremely favorable to the slow course of geochemical reactions.

The most important factor determining the possibility of the formation of iron ore deposits of the weathered crust is the climate, which will to a considerable degree determine the corresponding distribution of such deposits. The formation of hypergene iron ore deposits in general, and of weathering deposits in particu-lar, takes place intensively in regions charac-terized by a moist climate with moderate to high temperatures. For this reason the dis-tribution of iron-ore deposits of this genetic type will in some degree coincide with the boundaries of climatic zones. Thus emplace-ment of iron-ore deposits of the weathering crust will be determined by a favorable com-bination of the above-mentioned geotectonic, geomorphologic and physical geographic (namely climatic) conditions, and will also depend on the nature of the substrate being weathered.

It must also be remembered that many iron-ore deposits genetically associated with ancient weathered crusts, have been destroyed in varying degree by subsequent erosion. Only those deposits which were covered by trans-gressive sediments and buried under an over-burden of younger deposits have been preserved. Moreover it may be that there has been partial erosion of the weathering crust with a redeposi-tion of the material in it - that is, a formation of sedimentary deposits. For this reason it may be said more or less definitely that iron-ore deposits of this genetic type were formed no earlier than the Mesozoic. The oldest reliable remains of such weathering deposits

of iron ores are Jurassic (in the Urals).

Quaternary and Recent iron-ore deposits of weathering origin are distributed mainly throughout tropical and subtropical areas, where red soils and laterites are developed extensively. Such are the deposits in the Amazon River basin in South America, in parts of Africa and India, the Malay Archipelago, Australia, and the Philippines.

Principal examples of iron-ore metallogenic provinces whose deposits derive from an ancient weathering crust include the Northern Kazakhstan-Southern Urals Early Mesozoic province, the West Indies Tertiary province (Cuba and other islands), the Central African and the Indian Tertiary provinces.

It remains for us to consider the laws governing the geographic distribution or iron-ore deposits of metamorphic origin. These deposits are in the overwhelming majority Precambrian occurring in metamorphic rocks (mainly ferruginous quartzites, jaspilites and itabirites, and, more rarely, schists). These circumstances relate to a considerable degree the association of such deposits with basements of ancient platforms (shields). The map (fig. 1) shows clearly that metamorphic iron-ore deposits are almost exclusively confined to ancient platforms.

Iron ores of this genesis and age are concentrated in North America near Lake Superior and Lake Michigan, in Ontario, Quebec, and their vicinity; in South America (the Brazilian shield or platform) mainly in Brazil and adjacent Venezuela and Uruguay; in Africa (the African shield or platform) in Northern and Southern Rhodesia, Angola, the Belgian Congo, French Equatorial Africa, Togo, Sierra Leone, Mauritania, Kenya and Italian Somaliland; in Australia (the Australian shield or platform) in Western and Southern Australia; in Asia (the Siberian platform, the Indian and the Chinese platforms) in the Eastern Sayan, the Baykal area and the Transbaykal in the U.S.S.R., in Manchuria and the central provinces of China, in Korea and India; in Europe (the Scandinavian shield) in the Soviet Union (Krivoy Rog, Novyy Oskol, the Kola Peninsula, Finland and Norway. In these territories one may accordingly speak of a Precambrian iron ore metallogenic province.

This universal connection between metamorphic iron ore deposits and Precambrian formations must be stressed. In fact there is no large area of Precambrian rocks, especially of Algonkian age, which lack iron ores in considerable quantities. The total amount of Precambrian iron ores throughout the world is extremely large, being, according to N. M. Strakhov as much as three and a half trillion metric tons.

The laws governing the distribution of iron

ores within ancient platforms (shields) have as yet been very little described in the literature. The problem of the primary genesis of deposits of these ores has not been thoroughly studied [in the Soviet Union?] until recently. The most likely hypothesis, at least for some of these iron-ore deposits, is that of primary sedimentary formation in regions of Precambrian geosynclinal subsidence; moreover an important role in the formation of the ores was apparently played by submarine volcanic eruptions.

Within the Precambrian rocks in Europe two iron ore deposits have been discovered: one in Archean – the leptitic formation – and the other in the Proterozoic – the Saksagan series; one may accordingly speak of two Precambrian metallogenic epochs.

N. M. Strakhov (1947) believes that the epochs of iron-ore accumulation in the Precambrian were closely associated with geanticlinal regimes, and that the concentration of the iron on platforms was of secondary importance, and thus that the iron-ore deposits of the Precambrian are typical geosynclinal formations. S. P. Rodionov, while holding to this view, notes some definite regularities in the distribution of iron ore formations in the Precambrian rocks of the Ukranian SSR. According to S. P. Rodionov (1954), the remains of four submeridional folded structures composed of shale and hornfels, which make up the Saksagan series, can be traced from an area somewhat west of Krivoy Rog to the folded structures of the Orekhovo-Pavlograd zone in the east, the southern continuation of which is the ferruginous quartzite formation of Korsak-Mogila. The width of this ancient geosynclinal region of the Proterozoic is more than 200 km. Rodionov observes that the facies profile of the deposits in the Saksagan series indicates subsidence of a geosynclinal zone. He suggests that there is a connection between these above-mentioned Proterozoic folded structures and similar folded structures in the ore-bearing crystalline schists of Kursk and Karelia.

It is very likely that similar laws will govern the distribution of Precambrian iron-ore formations on other ancient platforms as well (such as the Canadian shield and the Brazilian platform), but, as already noted, these problems have still been little studied.

THE REGIONAL GEOLOGIC
DISTRIBUTION OF
MANGANESE-ORE DEPOSITS

Manganese is very close in its chemical properties to iron; it is not merely by chance that M. Goldshmidt, A. Ye. Fersman and others have assigned it to the geochemical family that includes iron. These two metals frequently behave in the same way in natural processes,

and this accounts for their simultaneous occurrence in certain types of deposits. But some types of deposits are known in which they do not form concentrations together. An especially indicative factor in this regard is the sharp difference between the behavior of iron and manganese in deep-seated processes of ore formation. In contrast to iron, which forms several types of deposits of igneous origin proper and is widespread in skarns and various hydrothermal deposits, manganese is not found in ore concentrations of igneous origin. Among postmagmatic deposits of manganese ores one may note only a comparatively small number of skarn and hydrothermal metasomatic deposits. But the distribution of both iron and manganese hydrothermal metasomatic ore deposits is very similar.

Mention was made earlier of the exclusive concentration and the association of siderite hydrothermal and metasomatic deposits in regions where Late Paleozoic and especially Mesozoic and Cenozoic folding occur; in southwest Europe, in the Balkans, in North Africa and adjoining regions of Asia Minor (the Mediterranean iron-ore metallogenic province). Such regularity is also observed in the distribution of hydrothermal manganese ore deposits, specifically: in their association with regions in which Alpine folding and igneous activity are developed in southwest Europe, in the Balkans and in North Africa (the Mediterranean manganese-ore metallogenic province). Moreover hydrothermal deposits of both iron and manganese ores are encountered separately.

Other regions in which hydrothermal manganese ore deposits occur are the zones of Mesozoic folding on the Pacific Ocean coast of North and South America, and also regions of Tertiary folding and volcanic activity in Japan.

Of special interest are the hydrothermal manganese-ore deposits rich in rhodochrosite in the Mediterranean metallogenic province, where they are associated with Recent igneous activity. There is basis for supposing that Recent igneous intrusives and extrusives are characterized by a considerably increased concentration of manganese; one part of the latter was brought in from the magma by gaseous and aqueous solutions and the other, in the form of admixtures, was found in silicates of extrusive, and especially effusive, rocks. Manganese enrichment appeared not only in the extensive development of manganese carbonates in various hydrothermal veins, which are commonly rich in polymetallic ores, but also in skarn deposits. For example, in Bulgaria, according to B. D. Bogdanov, such skarns consist primarily of rhodonite, johansenite and other manganese silicates.

Finally, manganese ore concentrations formed as a result of skarn and hydrothermal processes are definitely genetically associated with deposits of manganese ores of volcanic-sedimentary origin. It is supposed that the source material in the formation of the latter was manganese carried from the magma chamber together with silica and other components by means of hot mineral bottom sources acting together with submarine volcanic eruptions at the bottom of marine basins. But perhaps, and this is no less likely, in the formation of the so-called volcanogenic-sedimentary deposits the manganese was derived from pyroclastic accumulations rich in this metal, which were formed along with thick flows of andesite-dacite lavas in the Mesozoic and Cenozoic. Such pyroclastic products (various volcanic tuffs) which were only slightly consolidated, could, immediately after their ejection from the volcano's crater, be easily washed away and the manganese salts contained in them, being highly soluble, readily leached out.

It is interesting that sedimentary and volcanogenic-sedimentary manganese-ore deposits of Mesozoic and Cenozoic age coincide areally, in general, with the hydrothermal manganese-ore deposits of the single Mediterranean manganese-ore metallogenic province.

Let us now consider laws governing the emplacement of manganese-ore deposits of sedimentary origin; this, as is known, is the origin of the largest industrial deposits.

A study of the geochemistry, mineralogy, geology and metallogeny of manganese and manganese-ore deposits has occupied very many researchers, for example: V. I. Vernadskiy, A. G. Betekhtin, N. M. Strakhov, and N. S. Shatskiy.

In contrast to deposits of deep origin, hypergene deposits of manganese ores in regard to their regional geologic distribution, have much in common with hypergene iron-ore deposits. N. M. Strakhov (1947) considers that the overwhelming majority of manganese ores form an organic part of the iron-ore metallogenic provinces, combining with the iron ores in various complex ways, and also that the manganese-ore epochs are, in general, very close to the iron-ore epochs.

The metallogenic manganese epochs proposed by A. G. Betekhtin (1946) and by N. M. Strakhov (1947) are very similar; some of the differences between them are due to the fact that N. M. Strakhov's scheme encompasses the world, whereas that of Betekhtin is confined mainly to the Soviet Union.

Recently a great role in the formation of manganese sedimentary deposits has been attributed to volcanic processes. N. M. Strakhov (1947) very early noted the association of the southern Urals manganese deposits with

silicate extrusive series of the Devonian. He also pointed out the sedimentary-extrusive origin of the Early Carboniferous manganese deposits of Central Germany, in which he found a close association (both in space and stratigraphically) with submarine diabase extrusives. He also indicated close local connection in the Jurassic sedimentary manganese deposits, the primary ores of which, in the form of carbonate lenses and interbeds, were enclosed in the silicate-jasper series of California, the Alps, Malaya and elsewhere. At the same time, considering the Paleogene manganese province of Eurasia as a whole, Strakhov indicates that this is characterized by an absence of extrusive rocks syngenetic with the ores, and he considers that the paragenesis of the Paleogene manganese ores is purely sedimentary. It is impossible to agree with these views, however, as the Mediterranean basin is known to contain sedimentary deposits of manganese ores of Tertiary age, which are evidently formations genetically associated with volcanic processes.

Later, M. P. Kheraskov (1951) attempted to show the volcanogenic-sedimentary origin of the eastern Bashkirian manganese deposits contained in volcanogenic series of Upper Silurian and Devonian. Many of the arguments cited by him in favor of such a genesis of the eastern Bashkirian manganese deposits are quite convincing. But his broader general conclusions of the widespread distribution of manganese deposits of this genetic type are still unproven, and must therefore be considered only tentative. A local physical relationship between manganese deposits and extrusive rocks is not proof of bedded genetic relationship. For example, on the island of San Pietro in Italy and on the islands of the Aegean Sea in Greece, beds of manganese ores occur on the eroded surface of volcanic rocks and are also overlain by volcanics. Under these conditions it is not impossible that the manganese ores were formed by a leaching of the manganese from these volcanic rocks; the manganese can also have been taken from the latter during their weathering and in the period between the formation of the two volcanic flows. It is also impossible to agree with M. P. Kheraskov that if one admits a volcanic origin of the metal, the mechanism and the chemistry of the processes in the formation of the sedimentary manganese deposits become completely clear. In both interpretations of the genesis, these problems are still in many respects unclear and require further investigation. There is still also insufficient proof of the assertion that the jaspers were formed in close genetic association with submarine volcanic extrusions. It is quite possible that the jaspers are polygenetic formations, in which case the problem of the formation of the manganese deposits also becomes more complicated.

N. S. Shatskiy (1954) distinguishes a large group of manganese deposits of the volcano-genic-siliceous type, which he subdivides into several subgroups. Among these the most typical, from the standpoint of their genetic connection with volcanic eruptions and with siliceous and jasper rocks, are the manganese deposits in greenstone formations (spilitic-keratophyric) and the manganese deposits in jasper formations.

Shatskiy's ideas on the tectonic laws governing the distribution of the manganese formations distinguished by him are of great interest, but it is impossible to agree with all of this assertions. For example, he assumes a genetic connection between the manganese deposits with spilites and ophiolites, whereas one geochemical peculiarity of manganese (which distinguishes it from iron) is a tendency toward a greater concentration in acidic and alkaline magmas, and not in basic, much less ultrabasic magmas. Moreover, in distinguishing (among the volcanogenic-sedimentary deposits) between manganese deposits in the group of volcanogenic-siliceous formations of the first order (in particular, in greenstone and spilitic-keratophyric formations) and manganese deposits in the group of volcanogenic-siliceous formations of the second order (associated with porphyritic formations), Shatskiy observes the differences in the geologic positions of these groups of formations and deposits. In this way he destroys the fairly well-established concept of a single unified magmatic complex, a spilitic-keratophyric formation. But this can scarcely be done except on very solid grounds, especially in view of the complex phenomenon of differentiation that occurs in magma chambers.

The map of the world's manganese formations (fig. 2) shows both purely sedimentary and volcanogenic-sedimentary manganese deposits; but in view of what has been said above, the latter are distinguished only tentatively. Among the sedimentary manganese deposits on this map, several age subdivisions have been noted, reflecting periods of the most intensive formation of such deposits. These chronological subdivisions are based on the schemes proposed by A. G. Betekhtin and N. M. Strakhov. Among such periods or epochs one may distinguish the following:

1) Early Paleozoic (Cambrian and Ordovician;

2) Middle Paleozoic (Lower Devonian and Upper Devonian);

3) Late Paleozoic (Upper Carboniferous and Permian);

4) Mesozoic (Cretaceous and Jurassic);

5) Tertiary;

6) Recent.

In this categorization one may thus note a number of metallogenic provinces of sedimentary manganese ores, the majority of such provinces in early Paleozoic time being associated primarily with the margins of Precambrian platforms:

1) The North American metallogenic manganese province, which forms a belt along the north and southeast of the Canadian shield;

2) The South African metallogenic manganese province, fringing the African shield at the south;

3) The East Australian metallogenic manganese province, fringing the Australian Precambrian platform on the east;

4) The Northwest European metallogenic manganese province which adjoins the Scandinavian shield on the southwest;

5) The East Chinese metallogenic manganese province, and

6) The West Siberian metallogenic manganese province.

This disposition of the above-mentioned metallogenic provinces of early Paleozoic age, in relation to ancient platforms, to a certain degree determines both the geologic circumstances of the formation of such deposits, and possibly also the source of the metal that was removed by erosion from the rocks of the ancient platforms. Moreover in these metallogenic provinces one may observe deposits of manganese ores which are both purely sedimentary and volcanogenic-sedimentary in origin, bearing witness to their close genetic interrelationships.

In middle Paleozoic time one may indicate two metallogenic manganese provinces: 1) The Urals-Kazakhstan and 2) the West European. Both these provinces are characterized by extensive manifestations of the processes involved in the formation of manganese ores of volcanic-sedimentary and purely sedimentary origin. This metallogenic epoch, no less than that of the early Paleozoic, was apparently characterized by the formation of volcanic-sedimentary manganese ores.

A sharp contrast appears in the late Paleozoic manganese metallogenic epoch, during which purely sedimentary manganese deposits were formed, to the exclusion of every other type. During the late Paleozoic one may distinguish the following metallogenic manganese provinces:

1) the Southeast Chinese;
2) the Moroccan (in North Africa);
3) the Malay;
4) the West German, and
5) the Urals.

Sedimentary manganese deposits of the Mesozoic and Cenozoic metallogenic epochs are observed generally in the same provinces, which are concentrated in geosynclinal regions of the corresponding age. For example, in the Mesozoic one may observe two metallogenic provinces of sedimentary manganese ores:

1) the Mediterranean and

2) the Pacific Ocean.

The first of these encompasses North Africa, Italy, Czechoslovakia, the Balkan States, Asia Minor, the Caucasus and Central Asia, and is characterized primarily by the distribution of purely sedimentary (and to a considerably lesser degree of volcanic-sedimentary) manganese deposits.

The second province embraces a zone in which Mesozoic folding is developed on both the American continents along the coast of the Pacific Ocean (USA, Mexico, Chile). Primarily volcanogenic-sedimentary, and to a lesser extent purely sedimentary, manganese deposits are developed in this metallogenic province.

In the case of Tertiary times, mainly in the Paleogene metallogenic epoch, one may note two metallogenic provinces of sedimentary manganese ores, as in the Mesozoic: the Mediterranean and the Pacific Ocean.

The Mediterranean province extends from Spain in the Iberian Peninsula, and from Libya in North Africa, through France, Italy, Czechoslovakia, Bulgaria, Greece, Cyprus, the Ukraine, the Caucasus, Iran and farther eastward. This metallogenic province is represented not only by purely sedimentary but also by volcanic-sedimentary manganese deposits.

The Pacific Ocean province embraces the Pacific Ocean coast of the U. S. A., the countries of Central America (Cuba, Panama, Costa Rica) and the regions in which Tertiary formations occur in the Philippines, China and New Caledonia. The manganese deposits of this metallogenic province are also represented by both volcanogenic and purely sedimentary formations.

The Quaternary and Recent epoch in the metallogeny of manganese is characterized by the widespread development of lacustrine and swamp manganese ores. Manganese ores of this origin and age are particularly widespread in regions whose climate favored these processes. As examples, one may cite the North European metallogenic province, embracing the northwestern region of the Soviet Union, Finland, Sweden, Norway, Northern Germany and other areas; and the North American metallogenic province, encompassing Canada and the U. S. A.

Let us now consider the geologic laws

For delineation of Map Units I-V, see Figure 3

I   II   III   IV   V

Post-magmatic deposits:

▲ Bustamite-hausmannite-
   braunite in
   skarns

● Braunite-rhodochrosite-
   barite or quartz,
   frequently with sul-
   fides of lead, zinc,
   silver, copper, etc.

Deposits resulting from weathering:

◆ Opal-psilomelane-vernad-
   ite (caps of sediment-
   ary manganese deposits
   and manganiferous
   limestones)

◆ Psilomelane-vernadite
   (caps of metamorphogenic
   deposits)

◇ Hydrogoethite-psilomelane
   (commonly associated with
   karst in limestones)

Sedimentary deposits: opal-rhodochrosite-psilomelane-
pyrolusite within near-shore marine and lagoonal deposits:

■ Of Tertiary age

▨ Of Mesozoic age (mainly
   Jurassic and Cretaceous)

◪ Of late Paleozoic age
   (mainly of Late Carbon-
   iferous and Permian)

◩ Of middle Paleozoic age
   (mainly Late Devonian
   and Early Carboniferous)

⊠ Of early Paleozoic age
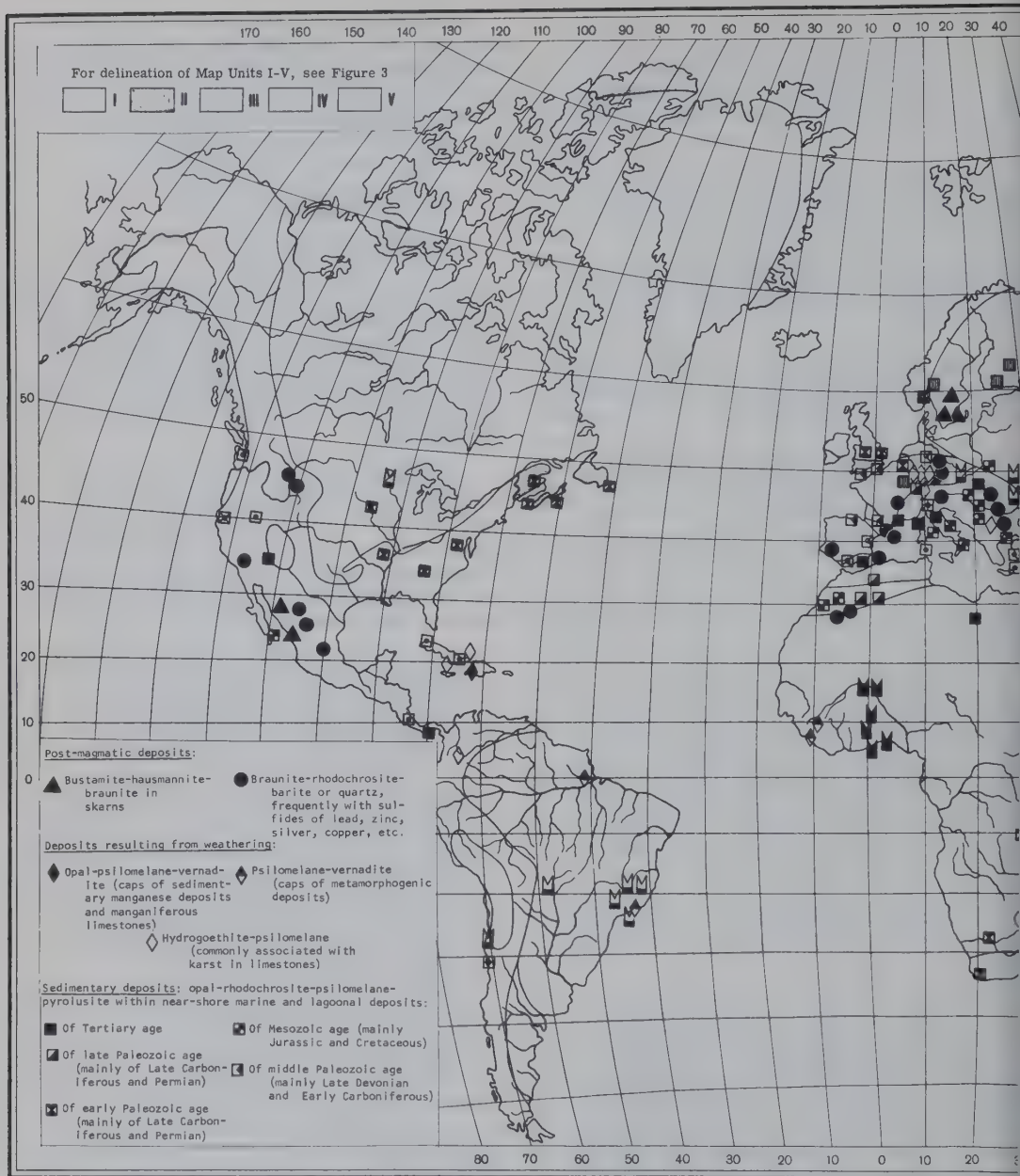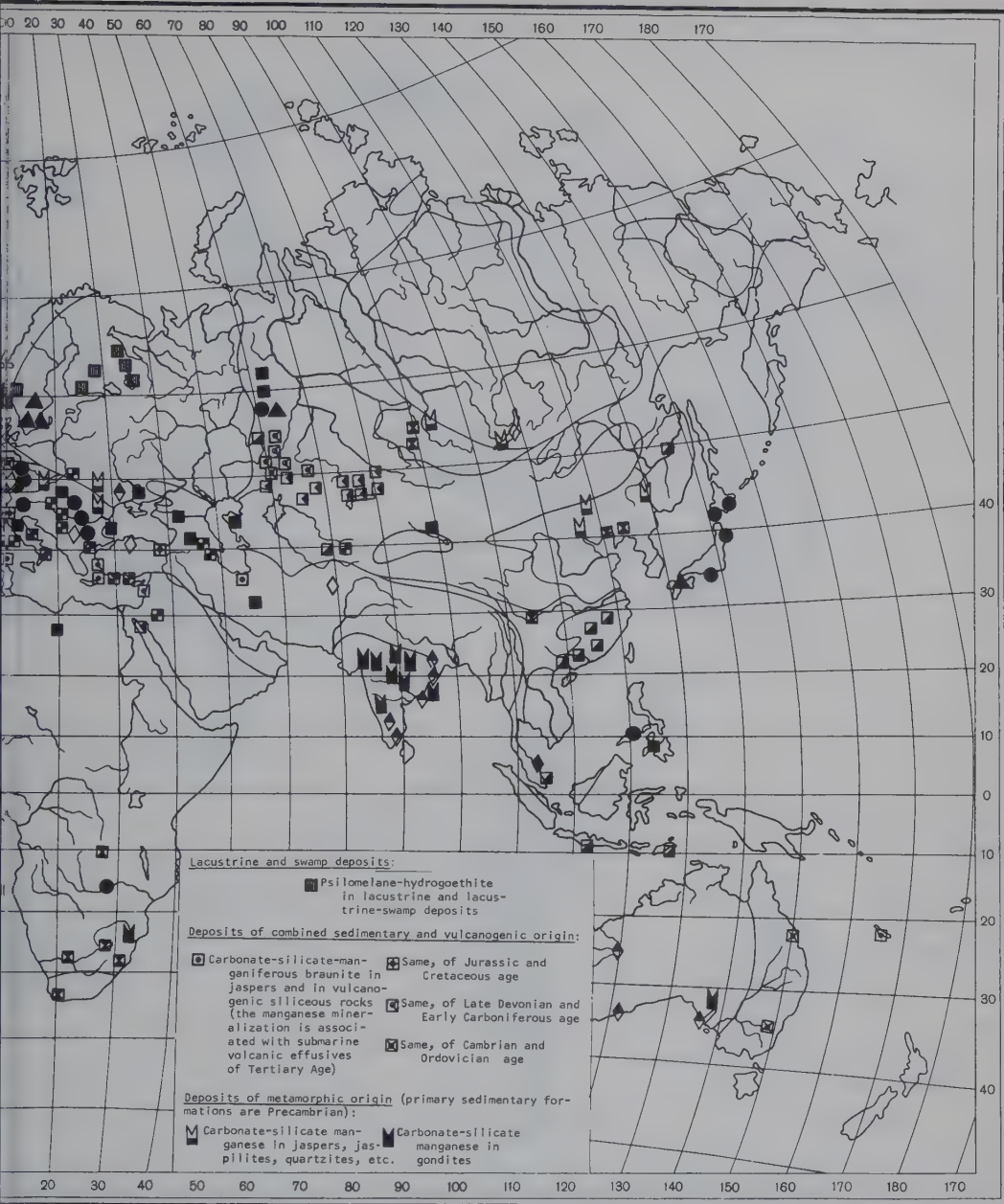   (mainly of Late Carbon-
   iferous and Permian)

FIGURE 2. Distribution of the

governing the distribution of the most ancient manganese ore formations, which are Precambrian. As mentioned before, they are for the most part represented by primary sedimentary strongly metamorphosed manganese-ore deposits. In many cases the degree of metamorphism will be so intensive that it is impossible to discover what the primary formations were. A. G. Betekhtin (1946) suggested classifying metamorphogenic manganese deposits into three groups, according to their degree of metamorphism. But certain difficulties arise

in applying this classification, because of the fact that the same deposit may contain ores with different degrees of metamorphism.

In determining the geologic laws governing the distribution of metamorphic manganese formations associated exclusively with Precambrian platforms, much interest attaches to the observations of N. S. Shatskiy (1954) on the genetic connection between the above-mentioned ores and the gondite and jaspilite formations, and to the fact that the latter metamorphic

Lacustrine and swamp deposits:

▣ Psilomelane-hydrogoethite in lacustrine and lacustrine-swamp deposits

Deposits of combined sedimentary and vulcanogenic origin:

◉ Carbonate-silicate-manganiferous braunite in jaspers and in vulcanogenic siliceous rocks (the manganese mineralization is associated with submarine volcanic effusives of Tertiary Age)

⊞ Same, of Jurassic and Cretaceous age

◧ Same, of Late Devonian and Early Carboniferous age

⊠ Same, of Cambrian and Ordovician age

Deposits of metamorphic origin (primary sedimentary formations are Precambrian):

M Carbonate-silicate manganese in jaspers, jaspilites, quartzites, etc.

M Carbonate-silicate manganese in gondites

lld's manganese-ore deposits

mplexes are the products of metamorphism rocks whose primary natures were essentially fferent. In particular, this applies to the pposition that the gondite formation and the nganese ores associated with it are products the metamorphism of volcanogenic-sedimenty spilite-keratophyric rocks and the manga se-bearing jaspers connected with them, thus presenting folded geosynclinal formations, ereas the jaspilite formation with the associ d manganese ores is the product of metamor sm of purely sedimentary rocks, whose

primary formation took place under different structural geologic conditions. N. S. Shatskiy's suppositions have still not been completely proved, but if confirmed, they will make it possible to explain the empirically discovered regularities in the distribution of metamorphogenic manganese ore formations.

In fact, as one may see from the map (fig. 2) the two genetic metamorphogenic types of manganese ore deposits have sharply different distributions. Deposits of silicate-manganese

Regions of Precambrian folding (ancient
platforms and shields I

Regions of early Paleozoic folding II

Regions of late Paleozoic folding III

Regions of Mesozoic folding IV

Regions of Cenozoic folding V

The definition and delineation of these Map Units
is the same for Figures 1 through 6

Deposits of igneous origin; chromite ores in
serpentinized dunites, peridotites, and serpentinites:

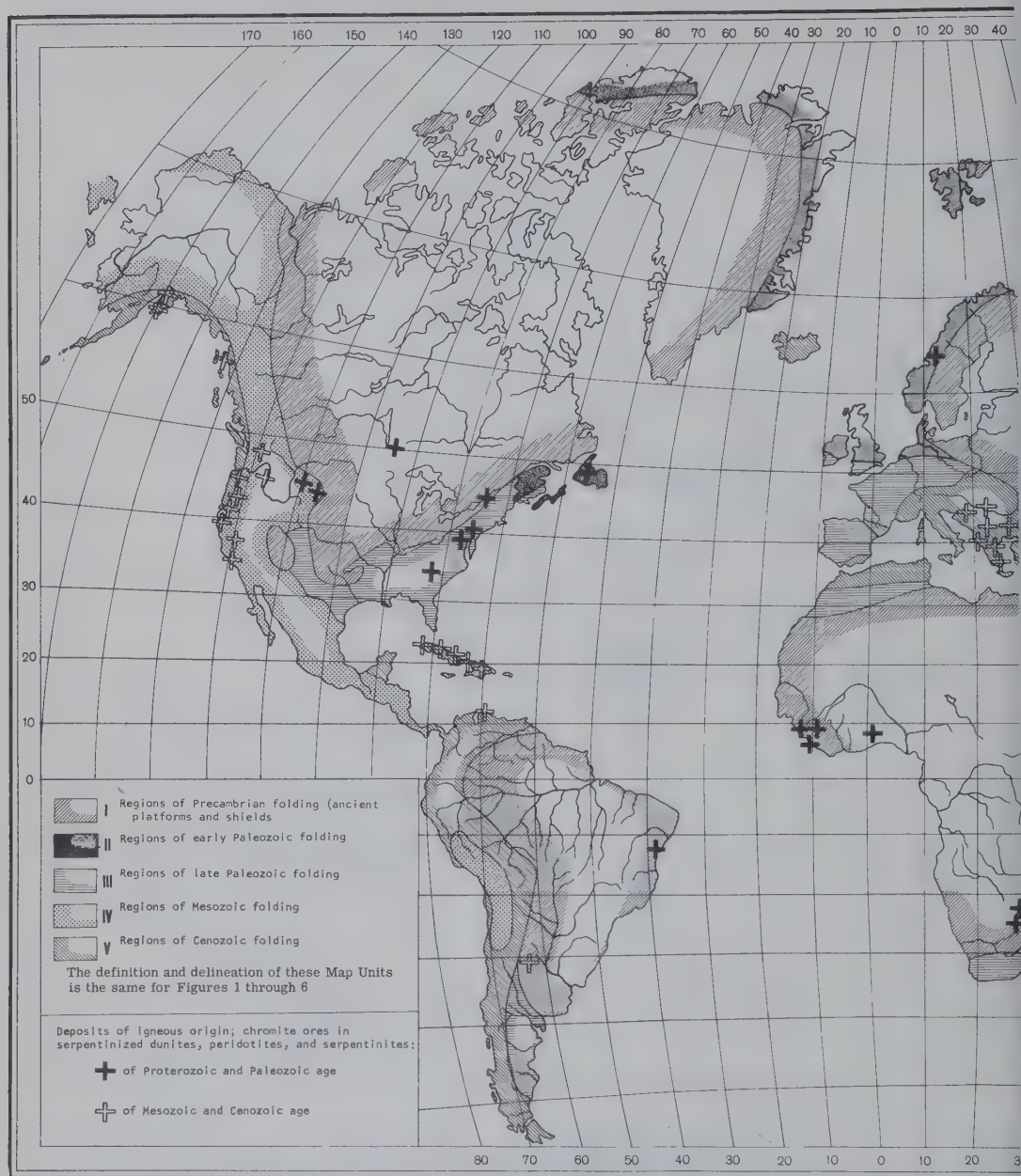of Proterozoic and Paleozoic age

of Mesozoic and Cenozoic age

FIGURE 3. Distribution of the

formations in gondites are observed only on the
Indian Precambrian platform and very locally
on the African Precambrian platform (Ghana),
and thus form only two metallogenic provinces.
Deposits of the carbonate-silicate-manganese
formation in jaspilites and siliceous schists
are widespread on the Canadian shield, the
Brazilian, Australian, Chinese and Siberian
Precambrian platforms and accordingly form
a number of manganese metallogenic provinces.

REGIONAL GEOLOGIC LAWS
GOVERNING THE DISTRIBUTION
OF CHROMIUM ORE DEPOSITS

The close geochemical connection between
chrome-spinel and ultrabasic igneous rock,
and thus the constant genetic association be-
tween chromite deposits and dunites, peridotites,
pyroxenites and the products of their altera-
tion -- serpentinites - is widely known. For
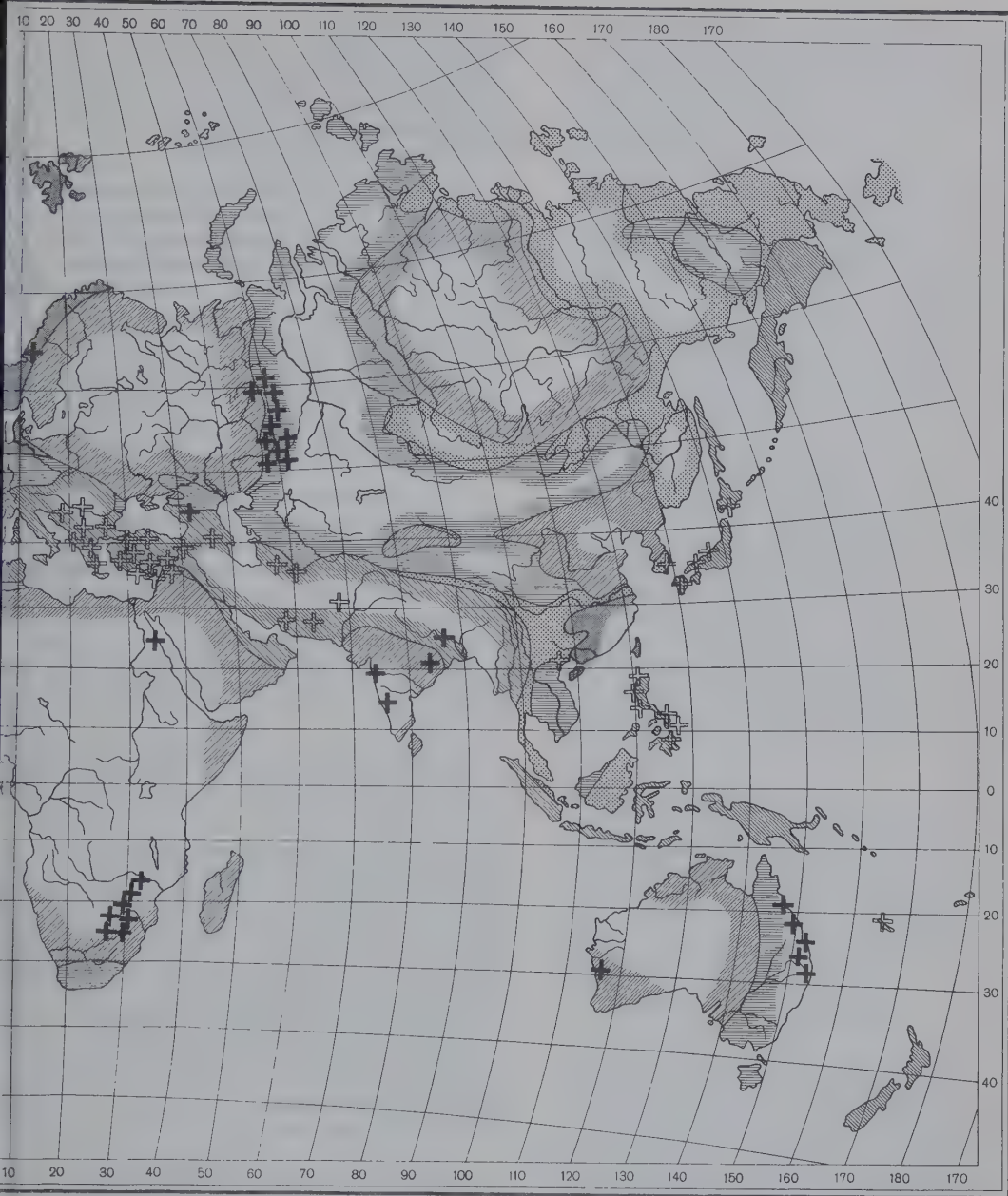
world's chromium-ore deposits

this reason the distribution of chromite deposits s to a considerable degree determined by laws overning the distribution of massifs of ultra-asic rocks. A search of the literature indi-ates that little attention has been devoted to hese questions. It is true that as early as he 1930's, V. N. Lodochnikov (1936) observed hat serpentinite massifs may have different ges – from Archean to Miocene. Of particu-arly great interest, according to V. N.

Lodochnikov, is the Mediterranean belt which extends almost parallel to the equator all the way from Spain to New Zealand, and with which many Tertiary serpentinite massifs are associated. Lodochnikov observed that this belt, also called the Eurasian-Australian belt, fringes the European and Asian continents on the south, and that the Tertiary serpen-tinite massifs everywhere occur in folded regions.

According to Lodochnikov, large massifs of serpentinized hyperbasites are also widespread in regions where Alpine folding is developed, on the Pacific coast of North America (U. S. A. and Canada). Finally, he pointed out the existence of enormous masses of hyperbasites in regions where more ancient folding is developed (Hercynian, Caledonian, etc.), in the Urals (Soviet Union), in the Appalachians (U. S. A.), in Eastern Australia and in other areas. These facts were the basis for Lodochnikov's assertion that hyperbasites and serpentinites are typical of folded regions, and that their intrusion took place during folding or shortly afterward; it is also characteristic that in folded regions serpentinites usually occur as bedded or layered deposits conformable with stratigraphically adjacent rocks.

On the map showing the distribution of the world's chromium-ore deposits (fig. 3) the striking feature is the different geologic positions of young and ancient chromium deposits. In the case of Mesozoic and Cenozoic chromium deposits one can observe the close connection of numerous and comparatively small chromium deposits primarily with zones of Tertiary folding, both in the Mediterranean geosynclinal zone and in the Pacific Ocean zone. The Mediterranean chromium metallogenic province includes deposits of chromium ores in Yugoslavia, Greece and Cyprus, Bulgaria, Turkey, the Caucasus and Iran.

The other province is that of the Pacific Ocean metallogenic zone. A large number of Mesozoic and Cenozoic chromium deposits are associated with the western half of the Pacific Ocean metallogenic zone (Japan, Korea, Eastern China, the Philippines and New Caledonia); the eastern half of this zone is characterized by an extensive development of Mesozoic and Cenozoic chromium deposits only in the northern part (the U. S. A. and Canada).

The largest metallogenic province of Paleozoic chromium deposits is that of the Urals, which includes scores of chromium deposits of Hercynian age, among which are the world's largest deposits: Kimpersay, Saranov and others. The ore bodies are lensed, commonly of enormous size. The other Hercynian metallogenic province is that of Eastern Australia. The chromium metallogenic province in the Appalachians (U. S. A.) is also Paleozoic.

Chromium deposits on ancient platforms are distinguished by a number of peculiar features. The ore bodies are in parallel layers conformable with layering of the ultrabasic rocks containing the ores; vein-like and irregular bodies are also observed. The ore bodies are commonly very large. Most are Precambrian, some Cambrian. There is a notable concentration of such deposits in the African shield, in South Africa (among them

one of the greatest deposits of the world, at Selukwe) and in West Africa (Sierra Leone and Togo), on the Canadian shield, on the Indian shield and other ancient platforms. One may accordingly speak of the South African, West African, Indian, North American and other chromium metallogenic provinces.

Among the deposits of nonferrous metal ores we shall consider only the most widespread, particularly: copper, lead-zinc and tin ores. The overwhelming majority from the genetic standpoint, are postmagmatic, primarily hydrothermal deposits. This is apparently the reason for the common occurrence of these deposits together. At the same time the distribution of each of these metals, and sometimes of the individual genetic types of their deposits, is determined by specific regional geologic laws individually typical.

A feature characteristic of the great majority of hydrothermal ore deposits of these metals is their association with regions of intensive folding in general, and of Mesozoic and Cenozoic folding in particular, and to a lesser degree of Paleozoic folding, whereas there is a relatively less widespread occurrence of igneous deposits proper (of copper), of metamorphogenic (of copper, lead and zinc) and of many pegmatite deposits (tin) associated with ancient platforms.

Above we have already noted the localization of a large number of hypogene iron-ore deposits and of numerous chromium deposits in a zone surrounding the Pacific Ocean on the one hand, and in a zone extending parallel to the equator from west to east, from the Mediterranean basin to the southeast of China, on the other hand; it is well known that these regions are characterized by intensive occurrence of Mesozoic and Cenozoic orogenic processes, igneous activity and associated mineralization. As regards the tin and copper ores, the first concept of the Pacific Ocean belt as a peculiar metallogenic province was suggested as early as the 1940's by S. S. Smirnov (1946). These ideas were later developed by Ye. A. Radkevich, V. K. Chaykovskiy and others.

Below we shall examine some of the laws governing the distribution of the principal genetic types of deposits of nonferrous metals.

REGIONAL GEOLOGIC LAWS
GOVERNING THE DISTRIBUTION
OF SILVER-LEAD-ZINC ORE DEPOSITS

In determining the position and association of postmagmatic deposits of silver-lead-zinc ores with the principal structural elements of the earth's crust, it may be stated that they are associated exclusively with regions of geosynclinal subsidence (fig. 4). These deposits are concentrated primarily in regions of Alpine, Hercynian and, to a lesser degree,

Caledonian folding, and where they are genetically associated with the corresponding occurrence of igneous activity. On platforms and shields, the regions of the most ancient folding, igneous activity occurs only in connection with deep faults. Here the scale on which the igneous activity occurs is very small in comparison with its manifestation in geosynclinal regions. For this reason, on ancient platforms (of pre-Paleozoic age) the ore deposits of deep-seated origin in general are not widespread; accordingly lead-zinc-silver ores also have a limited distribution in such regions.

A systematic analysis of the material on the geology of the world's principal silver-lead-zinc deposits of postmagmatic origin suggests that there are two essentially different groups. The first group consists of postmagmatic silver-lead-zinc deposits genetically connected with shallow intrusives occurring very close to the surface and with extrusive rocks of Mesozoic and Cenozoic age. The second group of postmagmatic silver-lead-zinc deposits is genetically associated with granodiorite intrusives primarily of Paleozoic age.

Let us examine in somewhat greater detail the geologic laws governing the distribution of these groups of silver-lead-zinc deposits and of their principal genetic types (ore formations). Deposits of Mesozoic and Cenozoic age are concentrated in regions of Alpine geosynclinal subsidence where folding has taken place. They make up several types of deposits.

1. Galena-sphalerite and galena-sphalerite-pyrite deposits in skarn carbonate rocks. Many of the very important deposits of these formations occur in the vicinity of Mesozoic and Cenozoic active intrusive massifs. Sometimes these deposits are found in skarns, the ore mineralization occurring much more commonly within the skarns, although under these circumstances there is generally a genetic connection between formation of the skarns and mineralization. Deposits of these ores exist in the Primorye region, in the Eastern Transbaykal, in Japan, China, the U.S.A. (Leadville, Park City, Tintic), in Argentina, Chile, Mexico, Italy (Cantiglia Maritima), Yugoslavia (Trepcha), Turkey and other countries.

2. Of considerable practical importance are the silver-galena-argentite-carbonate-quartz sulfo-salt deposits in propilitized Tertiary tuffs and lavas. Examples are the deposits in Mexico (Pauchuca, Guanahuato, Fresnillo), Chile, Peru, Honduras, Nicaragua, the U.S.A. (Comstock, Tonopah), Japan, Czechoslovakia, and Rumania.

The remaining ore deposits are less widely distributed. As one may see from the map (fig. 4), deposits of Mesozoic and Cenozoic age associated with regions of Alpine geosynclinal subsidence and folding are concentrated in two zones: a) one forming a circle along the shores of the Pacific Ocean on both continents of America and Asia, and b) extending from West to East, almost parallel to the equator, from the Iberian Peninsula and the extreme northwest of Africa through Southern Europe, Asia Minor, Central Asia to China.

These geostructural zones are characterized by a widespread and intensive development of volcanic processes, which began in the Mesozoic and ended mainly in the Tertiary, although in certain of the regions volcanic activity continued up to the present. Mention must also be made of the frequent shallow occurrence of the magma chambers, as indicated by observations in a number of places of gradual transitions from volcanic to hypabyssal facies of the eruptive rocks. The composition of the volcanogenic complex varies, but is represented mainly by andesites, dacites and rhyolites. The hypabyssal facies are represented by granites and granodiorites. In many areas of such deposits there is an intensive manifestation of igneous vein-rocks (quartz porphyries, granite porphyries, monzonite porphyries, etc.).

Many of the ore deposits of the group under consideration are associated not only genetically but also physically with volcanogenic rock: they are contained both in the pyroclastic products of volcanic eruptions, and in the lavas themselves. In the majority of these deposits one may observe intensive metamorphism of the ore-containing volcanogenic rock by ore-bearing postmagmatic gaseous solutions; here to a considerable degree there is also a manifestation of processes leading to the formation of propylites, alunites, kaolinites, zeolites, quartz, etc. There are very many occurrences of deposits of silver-lead-zinc ores in carbonate rocks, primarily in limestones.

The above-mentioned geologic conditions governing the formation of the deposits in this group to a considerable degree determine the morphological features of mineralization. The ores are mainly disseminated and vein-like bodies, numerous veins of small thickness and limited extent, which branch and frequently wedge out; finally, there are occurrences of dense sulfide ores of irregular shape associated with the gradual transition to disseminated ores and beyond these to oreless rocks.

The deposits in carbonate rocks are characterized by ore bodies resembling beds, mantles, columns, and other irregular shapes; disseminated ores and vein ores are also widespread. All these morpholgical features are to a definite degree determined by the metasomatism accompanying the formation of the ores.

We have already noted the extremely high concentration of silver-lead-zinc deposits of Mesozoic and Cenozoic age in two zones: one
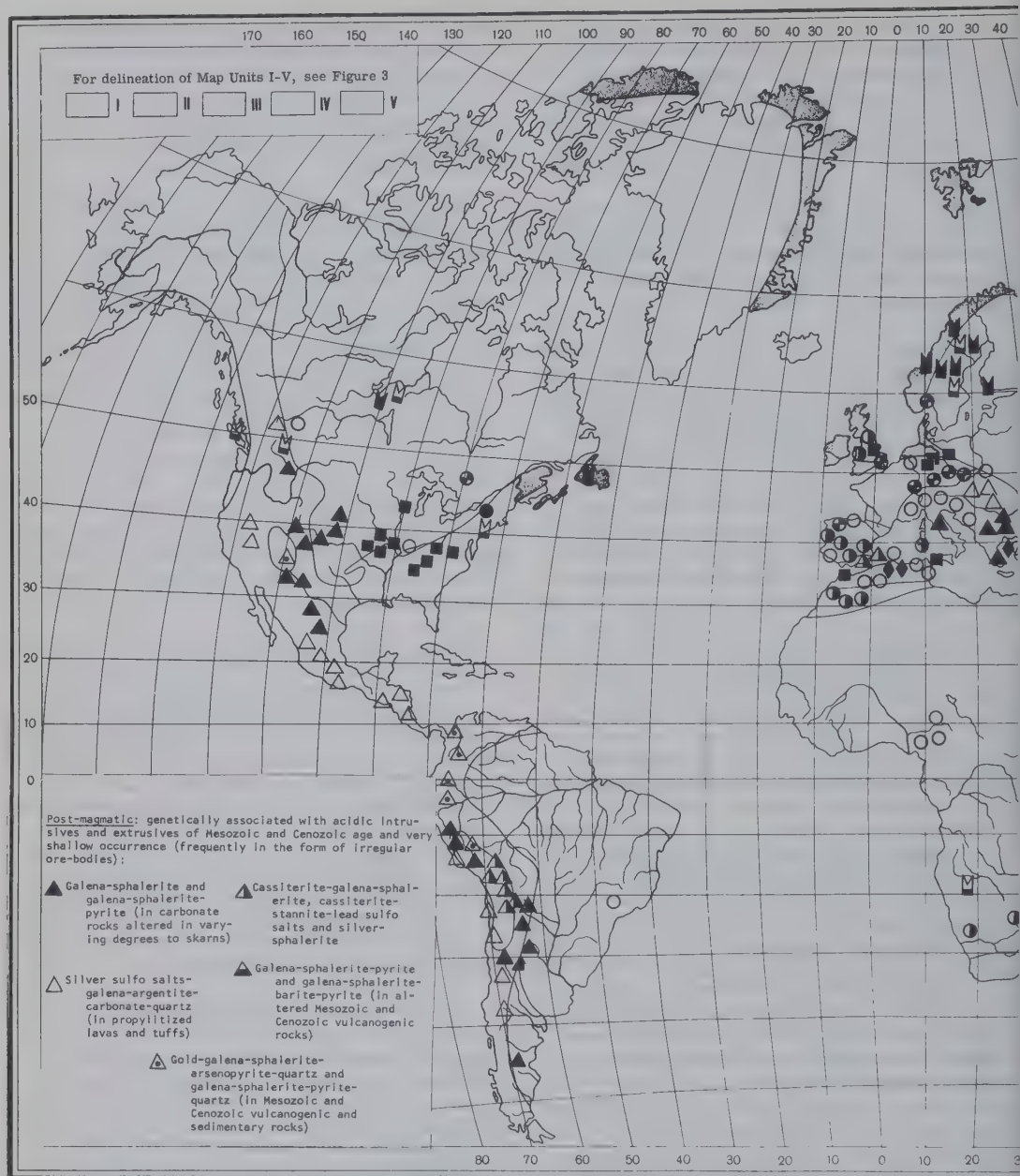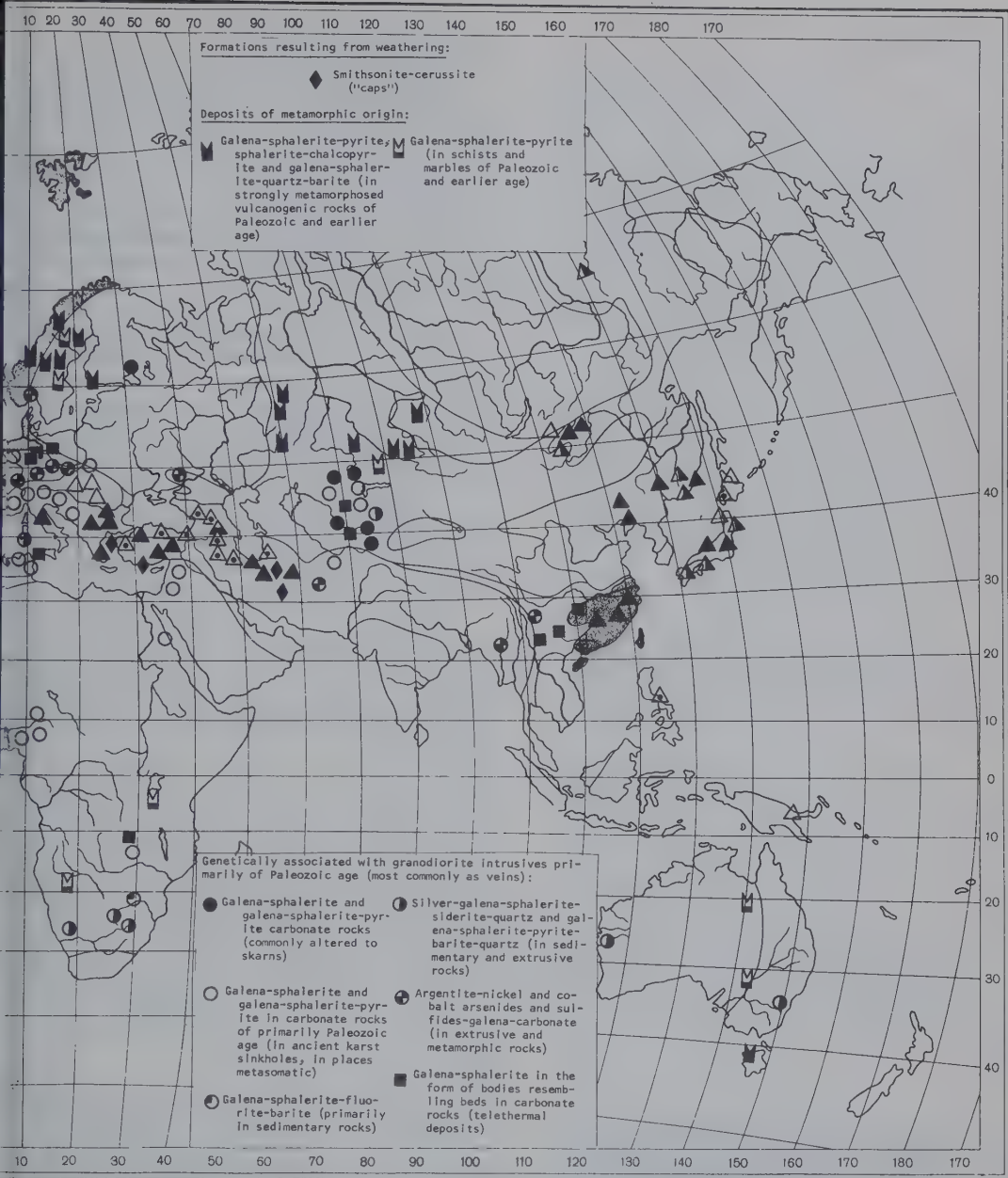
For delineation of Map Units I-V, see Figure 3
☐ I ☐ II ☐ III ☐ IV ☐ V

Post-magmatic: genetically associated with acidic intrusives and extrusives of Mesozoic and Cenozoic age and very shallow occurrence (frequently in the form of irregular ore-bodies):

▲ Galena-sphalerite and galena-sphalerite-pyrite (in carbonate rocks altered in varying degrees to skarns)

△ Cassiterite-galena-sphalerite, cassiterite-stannite-lead sulfo salts and silver-sphalerite

△ Silver sulfo salts-galena-argentite-carbonate-quartz (in propylitized lavas and tuffs)

△ Galena-sphalerite-pyrite and galena-sphalerite-barite-pyrite (in altered Mesozoic and Cenozoic vulcanogenic rocks)

△ Gold-galena-sphalerite-arsenopyrite-quartz and galena-sphalerite-pyrite-quartz (in Mesozoic and Cenozoic vulcanogenic and sedimentary rocks)

FIGURE 4. Distribution of the

surrounding the Pacific Ocean and the other extending parallel to the equator from the Mediterranean Sea basin in the west to southeast China in the east. We have also noted certain regularities in the distribution of these deposits within the above-mentioned zones. It must be stressed again that only the two above-mentioned genetic types of deposits have any extensive distribution: the galena-sphalerite and galena-sphalerite-pyrite deposits in carbonate rocks, and the sulfosalts of silver-galena-argentite-carbonate-quartz deposits in strongly

metamorphosed volcanic rocks. Within the Alpine geosyncline, deposits of both these types are associated with regions of the most intensive Mesozoic folding. At the same time one may observe certain regularities in the distribution of these two genetic types in relation to each other. In particular, on the American continent deposits of lead-silver ores, genetically and physically associated with centers of volcanic flows, are associated with the central parts of the zone of Mesozoic folding, whereas the belt of lead-zinc deposits in carbonate rocks is

Formations resulting from weathering:

♦ Smithsonite-cerussite
("caps")

Deposits of metamorphic origin:

⛏ Galena-sphalerite-pyrite,   ⛏ Galena-sphalerite-pyrite
sphalerite-chalcopyr-   (in schists and
ite and galena-sphaler-   marbles of Paleozoic
ite-quartz-barite (in   and earlier age)
strongly metamorphosed
vulcanogenic rocks of
Paleozoic and earlier
age)

Genetically associated with granodiorite intrusives pri-
marily of Paleozoic age (most commonly as veins):

● Galena-sphalerite and   ◐ Silver-galena-sphalerite-
galena-sphalerite-pyr-   siderite-quartz and gal-
ite carbonate rocks   ena-sphalerite-pyrite-
(commonly altered to   barite-quartz (in sedi-
skarns)   mentary and extrusive
                              rocks)

○ Galena-sphalerite and   ◑ Argentite-nickel and co-
galena-sphalerite-pyr-   balt arsenides and sul-
ite in carbonate rocks   fides-galena-carbonate
of primarily Paleozoic   (in extrusive and
age (in ancient karst   metamorphic rocks)
sinkholes, in places
metasomatic)   ■ Galena-sphalerite in the
                              form of bodies resemb-
◉ Galena-sphalerite-fluo-   ling beds in carbonate
rite-barite (primarily   rocks (telethermal
in sedimentary rocks)   deposits)

orld's silver, lead and zinc ore deposits

located mainly to the east of them, forming as it were the outer edge of the Pacific Ocean poly- metallic belt. The same regularity of distribu- tion may be traced, although less clearly, in the zone composing the circum-Pacific belt on the Asiatic continent. Finally, such a distribution of deposits of these two genetic types is also observed in the Mediterranean zone of the Alpine geosyncline. These regularities in the emplace- ment of the mineralization are apparently to be explained by the particular features of geotec- tonic and igneous development in these regions.

The other genetic types within the group of silver-lead-zinc deposits of Mesozoic and Cenozoic age generally have only a narrowly local distribution. In this regard it is worth paying some attention to the zone containing deposits of cassiterite-stannite-sphalerite-lead and silver sulfo-salts. Within the zone of post- magmatic silver-lead-zinc deposits of Mesozoic and Cenozoic age, stretching along the Pacific Ocean coast of the American continents, these tin-bearing deposits are characterized by an extremely high concentration in Bolivia, whereas

in the other countries of Latin America they are as a rule lacking.

We must stress certain differences between the mineral forms in which silver occurs in the silver-tin ores of Bolivia and the lead-silver ores of Mexico. In lead-silver ores, which do not contain tin, the silver is observed almost exclusively in the form of sulfo salts of silver and argentite, whereas in the silver-tin ores in Bolivia the silver is rarely found in the form of silver minerals as such, but much more frequently is chemically combined in the galena and in the sulfo-salts of lead and copper. The different paragenetic mineral associations in these formations, even though their geologic conditions of ore formation are very similar, indicates certain differences in the physiochemical and geochemical conditions of the ore-forming processes. The latter was apparently due to the genetic association between the ore-bearing gaseous solutions forming these two ore formations and the different magma differentiates, although this question has still not been sufficiently clarified with regard to the deposits under consideration here.

The postmagmatic silver-lead -zinc deposits of Paleozoic age are concentrated chiefly in regions of Hercynian and Caledonian folding. This group is represented by the following deposits:

1. Galena-sphalerite and galena sphalerite-pyrite deposits in carbonate rocks. These deposits are contained either directly in skarns, or else in limestones altered to marble. As examples one may cite certain deposits in Kazakhstan, Pitcorant, and Edwards Balmat (U.S.A.).

2. Silver-galena-sphalerite-siderite-quartz and galena-sphalerite-pyrite-barite-quartz deposits in sedimentary and eruptive rocks. These deposits are known in Central Asia, Morocco, the Union of South Africa, Australia and other places.

3. Argentite-nickel and cobalt arsenides and sulfides-galena-carbonate deposits in eruptive and metamorphic rocks. These deposits are encountered in the Black Forest, Andreasberg and Schneeberg (Germany), in Kongsberg (Norway), at Cobalt (Canada) and in other places.

4. Galena-sphalerite-fluorite-barite deposits occurring primarily in argillaceous and sandy shales. These deposits have been observed in Kazakhstan, the Union of South Africa and elsewhere, and are Paleozoic in age.

The number of deposits of this group is considerably smaller than that of the corresponding deposits of Mesozoic and Cenozoic age, due possibly to the fact that being older, they have been eroded to a greater degree. Genetically, the majority of deposits under consideration are associated with granodiorite intrusives, which are evidently mainly of Hercynian age. In certain cases these genetic connections and age relationships are only suppositional, but as a whole this group of postmagmatic silver-lead-zinc deposits differs essentially from the corresponding deposits of Mesozoic and Cenozoic age in its geotectonic occurrence, in the nature of the host rocks, in their contact-metamorphic aureoles around the ores, in their paragenetic mineral associations and in other important genetic indices.

Deposits of this group are characterized by their common association with large-scale faults and fault zones of Paleozoic age, which have ruptured the Paleozoic and pre-Paleozoic rocks, which have been subjected to some degree of dynamic metamorphism. The latter are represented by intrusives (granites, granodiorites, etc.) and more rarely by extrusives, in which case the extrusive rocks are usually somewhat older than the mineralization; the ore is sometimes also contained in various sedimentary and metamorphic rocks (gneisses, schists, etc.).

The ore-bearing granodiorite intrusive massifs of Paleozoic age are frequently worn down to the surface of the earth by erosion. On the peripheries of these intrusives one may sometimes observe lead-zinc mineralization, localized both in joints and faults. Apparently the minerals were formed by hot gaseous solutions separated from the magma in the deeper parts of these intrusive masses. In places the mineralization is connected with large faults and fault zones which dislocate both the intrusive masses and the complex of rocks overlying the given intrusive. In this case it must be assumed that the ore-bearing magma chamber was fairly deep. The ore bodies mainly have the form of veins, which frequently extend to great distances; in a number of deposits the number of ore bodies is very large, and their distribution follows definite regularities.

Everything that has been said indicates that there are great differences in the geotectonic conditions of formation of Paleozoic postmagmatic deposits and the corresponding Mesozoic and Cenozoic deposits. These differences are connected primarily with the geologic time of formation of the ore-bearing structural zones (Mesozoic and Cenozoic, and Paleozoic) and accordingly the time in which the magmatic processes took place; moreover the Paleozoic deposits are genetically associated primarily with comparatively deep granodiorite intrusives, in contrast to the Mesozoic and Cenozoic postmagmatic deposits, which are mainly connected with magma chambers of young volcanic activity. These differences in the geotectonic and igneous conditions of ore formation could not fail to have some effect on the physicochemical and geochemical conditions of these processes.

In contrast to the Mesozoic and Cenozoic deposits, where the genetic and physical connection with volcanic activity has determined the widespread occurrence of intensive alteration (propylitization, alunitization, etc.) of the ore-bearing volcanic rocks, in the case of the group of deposits considered here the alteration around the ore-bearing veins is usually not intensive. The wall rocks of the postmagmatic silver-lead-zinc deposits of Paleozoic age have usually undergone processes of quartz-formation, serpentinization, chloritization, and much more rarely skarn-formation and tourmalinization. These alterations occur only along the ore-bearing fractures, and to comparatively small distances from the ore bodies.

In regard to their material composition, the ores of the Paleozoic deposits and of the Mesozoic and Cenozoic deposits are generally very similar. This indicates a similarity in the physicochemical conditions of postmagmatic processes in the formation of the silver-lead-zinc deposits in different geologic periods. Such ore-forming minerals as galena and sphalerite are the principal minerals in the deposits under consideration. At the same time the genetic types of postmagmatic silver-lead-zinc deposits in the group are characterized by paragenetic mineral associations which are very similar but nevertheless peculiar to each type.

Among the lead-zinc deposits there is a large group whose genesis is still very controversial; this refers to the so-called "telethermal" deposits. Certain of them, at least, were evidently connected with processes of sedimentation in their primary formation. The scope of this article does not allow a discussion of these complex genetic problems, so that we must confine outselves to noting some of the peculiar geologic features of the regions in which these deposits are distributed. In North America, for example, deposits of this genetic type occur in a zone extending along the southern margin of the Canadian shield; in Central Europe the lead-zinc deposits form a narrow zone around the Scandinavian shield on the southwest; similar geologic conditions characterize the other deposits of this group as well.

The metamorphogenic deposits of silver-lead-zinc ores may be divided into at least two genetic types:

1) Galena-sphalerite-pyrite, sphalerite-chalcopyrite-pyrite and galena-sphalerite-quartz-barite deposits in strongly metamorphosed volcanogenic rocks of Paleozoic and older ages:

2) Galena-sphalerite-pyrite deposits in metamorphic slates and schists of Paleozoic and older age.

The origin of certain of these deposits is also still a matter for discussion. Nevertheless it is possible at the present time to make some observations regarding the regional geologic laws governing their distribution.

To begin with, it must be noted that deposits of the above-mentioned pyrite are fairly numerous. The deposits of the first type are genetically and physically associated with copper and copper-zinc pyrite metamorphogenic deposits and are contained in volcanogenic rocks, the products of thick, frequently submarine flows of diabase and albitophyres; they are commonly included in spilite-keratophyric rocks. These lava flows occurred mainly in areas of geosynclinal subsidence in early and middle Paleozoic times. This, in turn, has determined the distribution of the deposits of this group. The first ore mineralization was evidently genetically connected with comparatively shallow magma chambers. The primary ores were associated with tectonic zones of intensive shearing in the volcanogenic rocks and accompanied by very intensive processes of quartz formation, serpentinization, chloritization, pyritization and other processes in the hydrothermal alteration of the ore-bearing rock; processes of metasomatism were also widespread. The degree of later metamorphism to which these lead-zinc pyrite deposits were subjected varies within wide limits. Among the metallogenic provinces characterized by the development of such deposits are the Urals, the Altay, the Salair, the Scandinavian and others.

The deposits of the galena-sphalerite-pyrite formation in metamorphic slates and schists are primarily of Precambrian age (the second type of metamorphogenic deposits) and are associated almost exclusively with Precambrian platforms, which are of the same geologic age. These deposits occur in Australian and African Precambrian platforms and on the Canadian shield.

Thus the two genetic types of metamorphogenic galena-sphalerite-pyrite deposits are usually not encountered together, because those of the first type are associated with zones of Paleozoic folding in regions of older geosynclinal subsidence, and those of the second type with ancient Precambrian platforms.

REGIONAL GEOLOGIC LAWS GOVERNING THE DISTRIBUTION OF COPPER ORE DEPOSITS

In the case of the copper-ore formations, one may in general observe the same distribution of deposits as that of the silver-lead-zinc deposits, specifically: a great concentration of ores in regions where intensive Mesozoic and Cenozoic folding has taken place, both within the Mediterranean zone of the Alpine geosyn-
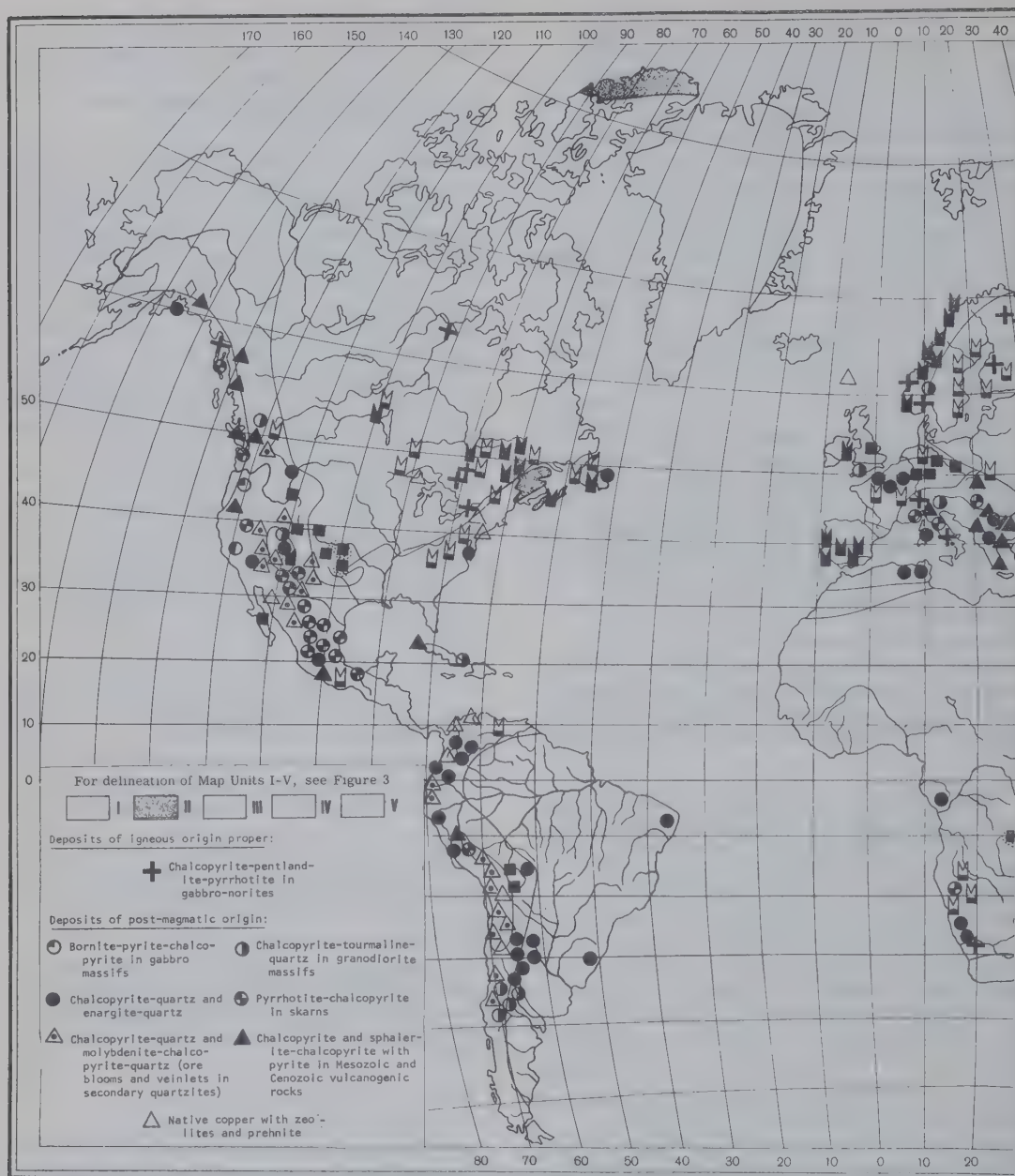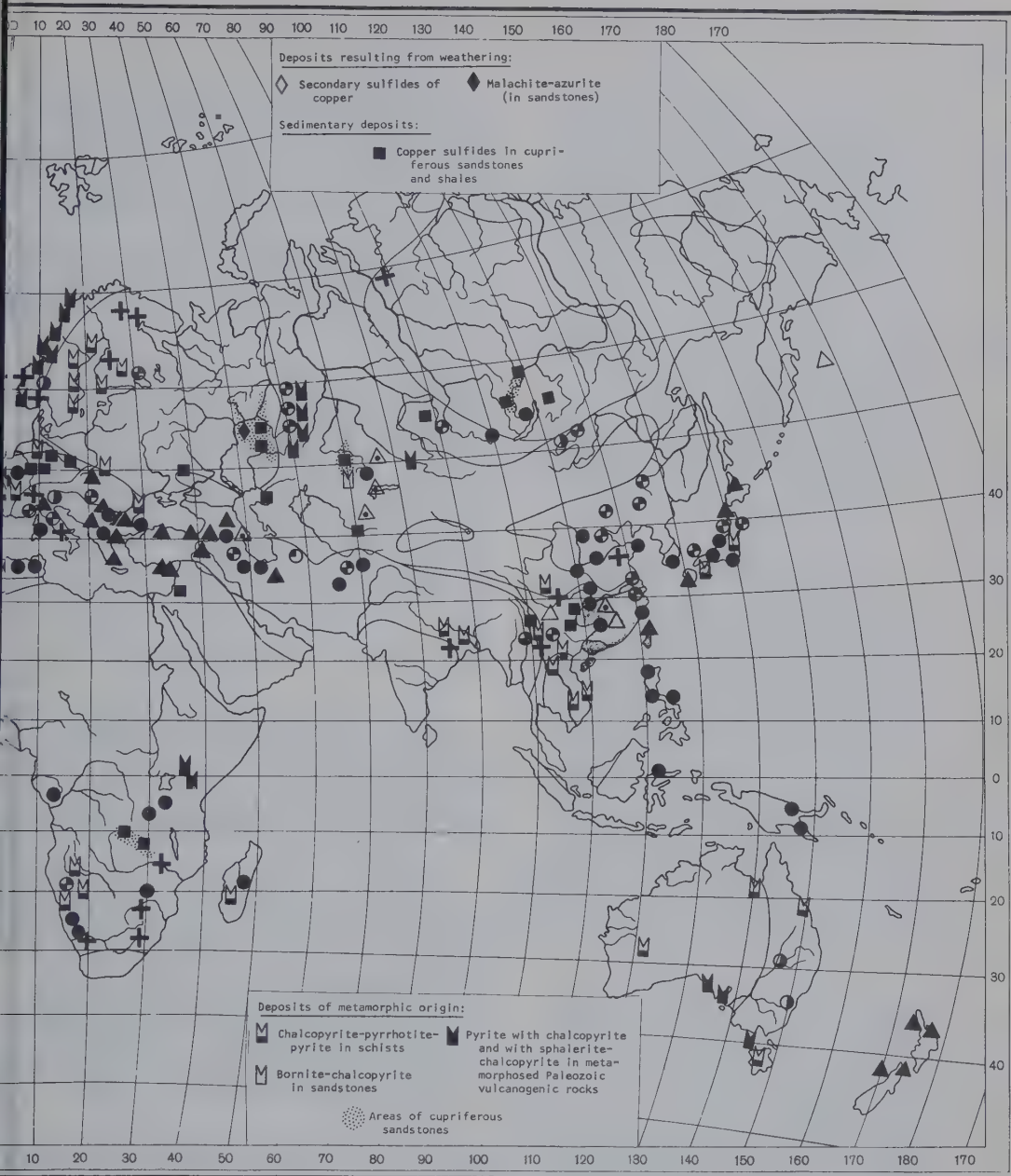
FIGURE 5. Distribution of the

cline and in the circum-Pacific belt. Since the overwhelming majority of the formations are of the same age, they are genetically associated with areas of especially intensive manifesta- tions of recent igneous activity. The degree of centration of the copper-ore deposits in the zones of Mesozoic folding is in general undoubt- edly higher than in regions of Paleozoic folding. The copper-ore deposits are apparently localized in regions of Mesozoic and Cenozoic folding and are accordingly connected with Tertiary and

Mesozoic igneous activity.

It is noteworthy that within the above-men- tioned regions one may observe a large number of varying genetic types of copper-ore deposits. As can be seen from the map (fig. 5), within the eastern half of the circum-Pacific metallo- genic zone there is a particularly large number of chalcopyrite-quartz deposits in secondary quartzites (disseminated and vein ores); there is a somewhat smaller number of pyrrhotite-

Deposits resulting from weathering:
◊ Secondary sulfides of copper
♦ Malachite-azurite (in sandstones)

Sedimentary deposits:
■ Copper sulfides in cupriferous sandstones and shales

Deposits of metamorphic origin:
Ⱶ Chalcopyrite-pyrrhotite-pyrite in schists
Ⱶ Bornite-chalcopyrite in sandstones
Ⱶ Pyrite with chalcopyrite and with sphalerite-chalcopyrite in metamorphosed Paleozoic vulcanogenic rocks
⁛ Areas of cupriferous sandstones

world's copper-ore deposits

chalcopyrite deposits in skarns, and a still smaller number of chalcopyrite-quartz vein formations and chalcopyrite-pyrite ores in volcanogenic Mesozoic and Cenozoic rocks. Apparently the majority of deposits of these ores are genetically associated with magma chambers occurring close to the surface. This explains the fact that many of these deposits (especially the disseminated copper ores in secondary quartzites) are associated with the extrusive facies of the magma; in this connection,

one may observe peculiar alterations in the ore-bearing rocks (tourmalinization, propylitization, alunitization, quartz formation etc.). The gradual transitions from hypabyssal to extrusive facies indicate that the ore-bearing magma chambers were close to the earth's surface.

It should be noted here that the majority of deposits of disseminated copper ores in secondary quartzites are apparently both

genetically and physically associated with zones of Mesozoic folding and igneous activity of the same age.

In the western half of the Pacific Ocean metallogenic belt, the same genetic types of copper ore deposits are developed (with the exception of the chalcopyrite-quartz formations in secondary quartzites). This may be due to the formation of deposits of this genetic type primarily in regions of Mesozoic folding and volcanic activity of the same age.

The very same metallogenic features as appear in the western half of the circum-Pacific belt are manifested in the Mediterranean zone of the Alpine geosyncline and the folded and igneous processes developed there - that is, a predominance of chalcopyrite and sphalerite-chalcopyrite pyritic formations in young volcano-genic rocks, pyrrhotite- chalcopyrite formations in skarns and vein chalcopyrite-quartz and enargite-quartz formation.

A somewhat different character has been observed in the copper mineralization of regions where Paleozoic folding and accompanying igneous activity appear. In such regions it is more typical to find a distribution of deposits of various pyrite formations pyritic chalcopyrite-pyrite, sphalerite-chalcopyrite-pyrite, etc.), often in close physical association with Paleozoic volcanogenic rock complexes. Deposits of the pyrrhotite-chalcopyrite formation in skarns and the vein chalcopyrite-quartz and chalcopyrite-tourmaline-quartz formations are less wide-spread. The latter two genetic types are distinguished from the analogous genetic types of Mesozoic and Cenozoic age by deeper occurrence of the metalliferous magma chamber (as may be judged by a combination of geologic and petrographic data).

Regarding the deposits of the pyrite formations, the majority of them are apparently typical primary medium-temperature hydrothermal formations with corresponding occurrence of the ore-bearing magma chambers. It is not impossible that the depth of occurrence of the latter in some regions was very small, so that in this case the processes of mineralization could have been of a fumarole nature: but such cases are apparently extremely rare, even for the Urals.

Deposits of the chalcopyrite-quartz formations in secondary quartzites are known to be widespread in Kazakhstan. According to N. I. Nakovnik (1954), the formation of these deposits is associated with extrusive processes during the late Paleozoic. Although this would agree with what has already been noted in the case of the genetically similar deposits of the Mesozoic in the eastern half of the circum-Pacific metallogenic belt, how definite this is in the case of Kazakhstan is a moot point, inasmuch

as typical manifestations of extrusive processes have not been observed in these deposits.

Copper mineralization of a peculiar nature has been found on ancient Precambrian platforms and shields. In the case of these structural elements of the earth's crust, the most characteristic genetic type of copper ores is igneous deposits proper, represented by chalcopyrite-pentlandite-pyrrhotite formations in gabbro-norites and pyroxenites. This is well illustrated by the localization of deposits of this formation in the Canadian and Scandinavian shields, as well as on the African, Siberian, Indian, Chinese and other platforms (fig. 5).

Another genetic type of deposit frequently observed on platforms and shields (the Canadian, Scandinavian, Australian, African and other shields), is that of pyrite deposits which may or may not contain nickel in the ores, occurring in series of metamorphic slates and schists. In the majority of these deposits, no clear connection has been established with any definite igneous intrusives. These are apparently hydrothermal formations, often high-temperature. Their frequent occurrence in the same regions as copper-nickel deposits in gabbro-norites suggests a possible relationship with the latter.

The connection between igneous deposits of copper-nickel ores associated with gabbro-norites and ancient platforms provides a basis for the supposition that ore-bearing, poorly differentiated gabbro-norite magmas were injected along deep-seated faults in the platforms.

The problem of sedimentary copper deposits is still a subject of controversy. Nevertheless, the majority of these deposits in sandstones and shales are sedimentary. Many of them occur on platforms close to the source of the clastic transported material. But one must also take account of the physical geographic conditions (particularly the climate). It is well known that the Permian period was in this respect most favorable for the concentration of copper ores.

The most ancient sedimentary deposits definitely bear traces of superimposed diagenesis and metamorphism.

REGIONAL TECTONIC DISTRIBUTION OF TIN ORE DEPOSITS

The regional distribution of tin ore deposits, like that of the deposits of copper and silver-lead-zinc ores, depends mainly on the appropriate combination of geotectonic, igneous, geochemical and physicochemical factors. The distribution of tin-ore placer deposits, which are also of great importance, depends considerably on physical geography.

S. S. Smirnov (1946), Ye. A. Radkevich (1947),

. D. Levitskiy and other Soviet geologists have discussed the laws governing the distribution of the tin ore deposits, both within the Soviet Union and other countries. These problems have also been discussed to some degree in publications by W. R. Johns, A. Cissard (?) and other non-Soviet authors. Below we shall consider some of the laws governing the physical emplacement of tin ore deposits.

Of particular interest in this regard are the papers by S. S. Smirnov (1946) on the circum-Pacific belt, by which he means the zone of Mesozoic and Cenozoic folding, igneous activity and metallogeny which surrounds the Pacific Ocean in an almost continuous ring. Smirnov stresses the fact that within this zone, against a background of weak ancient stages of metallogenesis, there is a striking manifestation of extremely rich Mesozoic and Cenozoic ore mineralization. Gold and silver, tin, tungsten and molybdenum, lead, zinc and copper, arsenic, bismuth, mercury and a number of other metals are encountered in the circum-Pacific belt in numerous and frequently very large concentrations, forming an almost continuous zone of "young" ore deposits.

Within the Pacific Ocean belt, Smirnov has tentatively distinguished two zones with somewhat different metallogenies - an inner zone close to the Ocean itself, and another zone which surrounds the first on the side of the continents. The inner zone is characterized by a widespread distribution of concentrations of copper and by the exclusive concentrations of tin. The outer zone, on the other hand, is poor in copper and extremely rich in tin; this zone encompasses almost all of the eastern margin of Asia and, in the form of separate areas, has been observed in North and South America.

We have already considered the geologic laws governing the distribution of copper ores and silver-lead-zinc deposits, and observed the connection between these deposits (genetically associated with Mesozoic and Cenozoic igneous activity) and zones of Alpine folding, developed both in the regions of the Mediterranean geosyncline (Tethys) and in the circum-Pacific belt.

In actual fact, an enormous number of different genetic types of copper ore and silver-lead-zinc deposits are closely associated with igneous processes in the late Mesozoic and Cenozoic eras, manifested in the zones of the Alpine orogenesis. Moreover these deposits form almost continuous zones along the shores of the Pacific Ocean on both the continents of America and Asia.

Nevertheless this definite regularity in the distribution of copper and silver-lead-zinc deposits in the circum-Pacific belt does not extend to the tin ore deposits, for which circumstances

there are definite reasons to be set forth below. This metallogenic nature of the Pacific Ocean belt was noted as early as 1946 by Ye. A. Radkevich, who indicated that the Pacific Ocean ore belt, as far as tin ore mineralization is concerned, does not have the significance attributed to it by S. S. Smirnov. In Radkevich's opinion the circum-Pacific ore belt has a complex structural geologic and metallogenic nature; it may be tentatively characterized as a region of intensified tin ore mineralization which encompasses different ore provinces associated with various zones; moreover the tin ore provinces make up only a small part (less than one-tenth) of the circum-Pacific belt.

Ye. A. Radkevich (1947) has established some other regularities in the distribution of tin ore deposits. For example, segregating the cassiterite-sulfide deposits as an independent group, she has indicated their usual association with zones of recent folding and noted that a typical peculiarity of such deposits is their formation at small depths far from ore-bearing intrusives.

In contrast to the cassiterite-sulfide deposits, veins of cassiterite-quartz formations are widespread within deeply eroded Precambrian folded formations and folded zones of Caledonian, Hercynian and Mesozoic age; in the folded zones of Tertiary age these deposits are not widespread. Ye. A. Radkevich has also noted that tin-bearing pegmatites are in the majority of cases encountered in deeply eroded regions of crystalline shields and ancient folded structures. The map of the world's tin ore formations accompanying this article (fig. 6) confirms the validity of many of the views of S. S. Smirnov and Ye. A. Radkevich, although there are also certain other regularities which were not indicated earlier or not sufficiently stressed.

The total number of tin ore deposits of Mesozoic and Cenozoic age known at the present time considerably exceeds the number of tin ore deposits of pre-Mesozoic age. The number of tin ore deposits on ancient shields and platforms, representing Precambrian folded structures, is small; this is apparently due to the deep erosion that has taken place in these regions. It is interesting that in the majority of cases these deposits consist of tin-bearing pegmatites. Deposits of cassiterite-quartz and tin-bearing greisens are somewhat smaller in number; skarn deposits of tin-ores, characterized by the presence of sulfides and in particular of chalcopyrite, are encountered much more rarely under these geologic conditions. Such ancient metallogenic epochs and provinces have been found in Africa (the African shield), Australia (the Australian shield), Brazil (the Brazilian shield), the U. S. A. and Canada (the Canadian shield) and elsewhere.
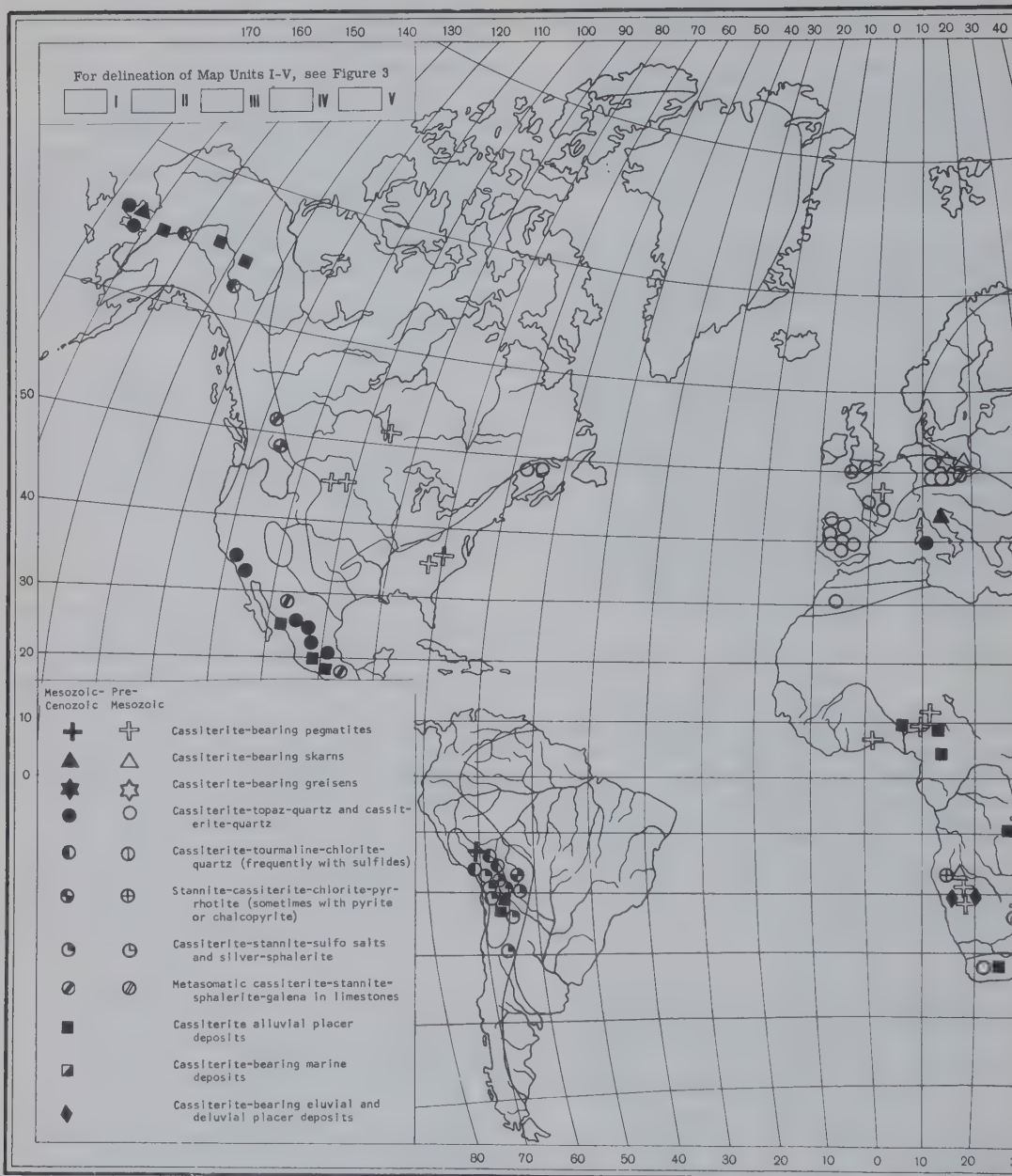
Regions of late Paleozoic folding are

For delineation of Map Units I-V, see Figure 3

☐ I ☐ II ☐ III ☐ IV ☐ V

Mesozoic-Cenozoic	Pre-Mesozoic	
✛	✣	Cassiterite-bearing pegmatites
▲	△	Cassiterite-bearing skarns
★	☆	Cassiterite-bearing greisens
●	○	Cassiterite-topaz-quartz and cassiterite-quartz
◑	⊕	Cassiterite-tourmaline-chlorite-quartz (frequently with sulfides)
◓	⊕	Stannite-cassiterite-chlorite-pyrrhotite (sometimes with pyrite or chalcopyrite)
◕	◔	Cassiterite-stannite-sulfo salts and silver-sphalerite
⊘	⊘	Metasomatic cassiterite-stannite-sphalerite-galena in limestones
■		Cassiterite alluvial placer deposits
◨		Cassiterite-bearing marine deposits
◆		Cassiterite-bearing eluvial and deluvial placer deposits

FIGURE 6.   Distribution of the

characterized by a wider distribution of tin ore deposits than Precambrian platforms, although fewer tin deposits are found than in regions of Mesozoic and Cenozoic folding.   In regions of late Paleozoic folding the most extensively distributed deposits are of the cassiterite-quartz type; considerably less widely distributed are tin-bearing pegmatites and tin-bearing greisens; but in certain regions there is a considerably greater occurrence of cassiterite-sulfide deposits (England, New South Wales in Australia, Tasmania).

The formation and corresponding distribution of cassiterite-sulfide deposits is apparently the result primarily of a favorable combination of geotectonic and magmatic factors.   One cannot explain the presence or absence of these deposits in a given region only by the depth of erosion, as suggested by Ye. A. Radkevich. It is quite clear that there is no particular relationship between the distribution of the individual genetic types of tin ore deposits and the depth of erosion.   One can speak only (as already noted) of a possible relationship
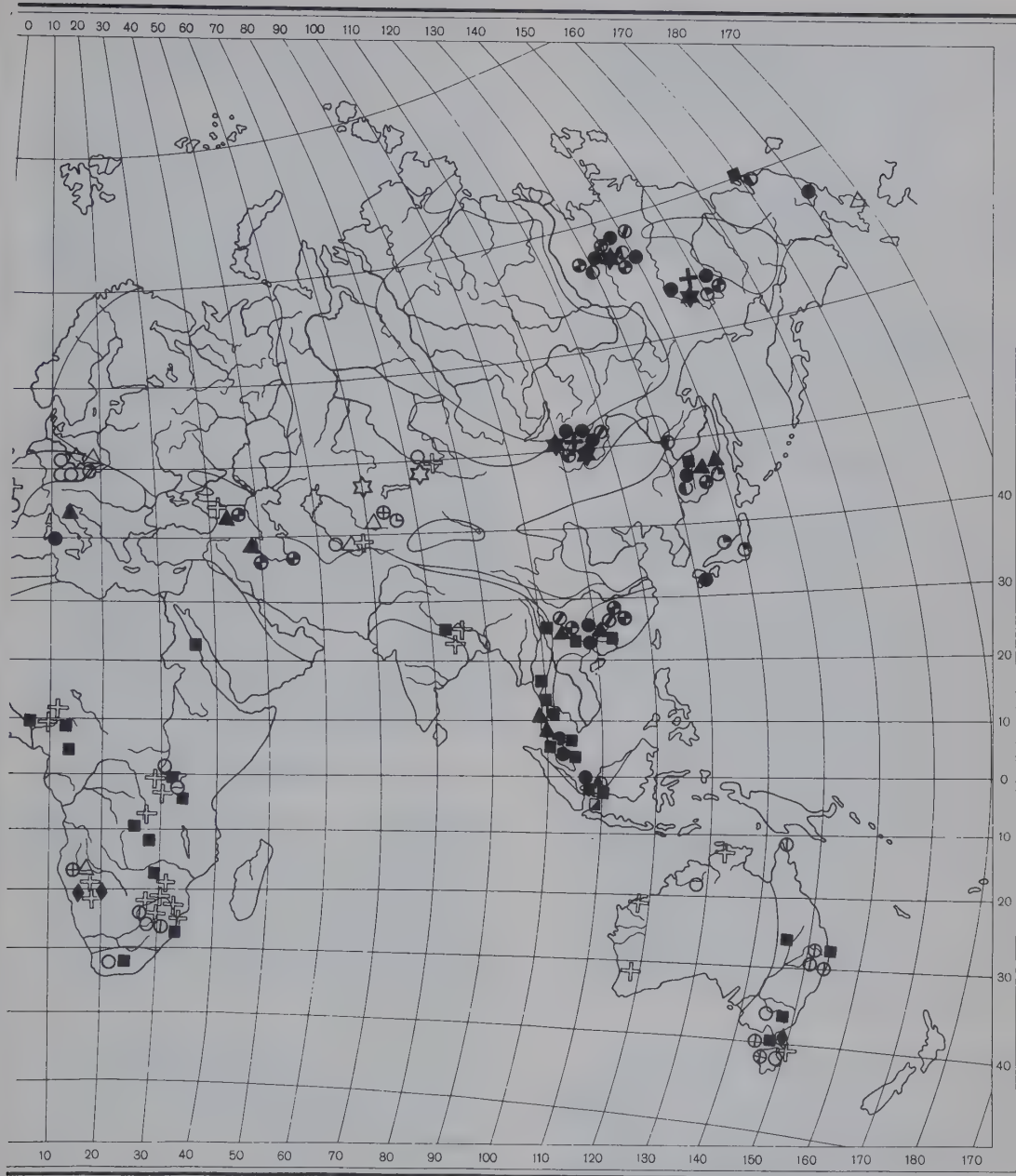
world's tin ore deposits

between the quantitative development of tin mineralization in general and the depth of erosion. Examples of late Paleozoic tin metallogenic provinces are found in western Europe and eastern Australian.

One of the most interesting and important regularities in the distribution of tin ore deposits of Mesozoic and Cenozoic age is their primary connection with regions where folding ended for the most part in the Mesozoic. Con-

sidering the time of occurrence of the igneous activity in these regions, there is sufficient basis for supposing that the greater part of the tin ore deposits in Mesozoic folded structures also are Mesozoic in time of formation. One must note the very slight manifestation of tin ore mineralization within Tertiary folded regions. The latter, it is well-known, are widespread in the Mediterranean zone of the Alpine geosyncline, these regions being characterized by an almost total absence of tin ores.

Thus the Mesozoic tin ore metallogenic epoch appears very clearly in the geologic history of the earth's crust; moreover during this time, as may be seen on the map (fig. 6), were formed the overwhelming majority of the tin ore formations existing at the present time, including those of the greatest industrial importance. Therefore the regional distribution of Mesozoic folding accordingly determines the distribution of the tin ore deposits within the Pacific Ocean ore belt and, in particular, the absence of a discontinuous belt of tin ore deposits.

In this respect the distribution of the tin ore formations differs essentially from that of the silver-lead-zinc ore formations, and also that of the copper ore formations (with the exception of the chalcopyrite deposits in secondary quartz-ites), for which such a close local association with regions of Mesozoic folding has not been noted.

All this amounts to one of the most important reasons for the absence, in the case of tin-ore formations, of the same regional geologic regularities in the distribution as were discovered and indicated for deposits of the silver-lead-zinc ore formation - a zone along the Pacific Ocean coast on the continents of America and Asia, and another zone extending almost parallel to the equator, from the Iberian Peninsula and the extreme northwest corner of Africa, through southern Europe, Asia Minor, and Central Asia to China.

Moving on to the distribution of the tin ore deposits within regions with a predominant development of Mesozoic folding, it must be noted that here one may encounter literally all the genetic types of tin ore deposits of deep-seated origin. As regards the cassiterite-sulfide formations, in all regions of Mesozoic folding these are considerably more widespread than in regions of older folding.

Tin ore deposits are not developed in all regions of Mesozoic folding; moreover, their distribution shows no concentration in belts or zones. On the other hand, they are concentrated in narrowly local, usually isolated, regions. Each of these regions is a geologic and metallo-genic province with a characteristically typical manifestation of orogenic, igneous and ore-forming processes. Within the regions of Mesozoic folding one may note no less than twelve or thirteen such metallogenic provinces, and within each of these provinces are almost all the genetic types of tin ore deposits. This peculiar metallogenic nature of the regions of Mesozoic folding is apparently due to the corre-sponding development of igneous processes in these metallogenic provinces.

Actually, the most acidic varieties of granites are apparently associated with ore-bearing pegmatites and greisens, and also the veins of the cassiterite-quartz formations; at the same time the cassiterite-sulfide deposits are associ-ated mainly with granodiorite facies of the magma in certain regions. The occurrence of the granodiorite magmas may be attributed to the peculiar course of the differentiation under the appropriate geotectonic circumstances. A large role was played by carbonate rocks, which were involved in the formation of various tin-bearing skarns. It is not impossible that there was assimilation, which would have facilitated the occurrence of the more basic magma facies and the iron-rich tin ore deposits genetically associated with them. An interesting aspect is the enrichment of these formations in boron as encountered in a number of minerals (tourma-line, datolite, nordenscheldite, pageite, etc.).

The local emplacement of the tin ore con-centrations was determined by the corresponding localization of the igneous activity in these provinces. Depending on the nature of the differentiation in the magma chamber, which, in turn, depend on the tectonic development of the geologic and metallogenic province, the various genetic types of tin ore deposits might be developed to a greater or lesser degree. The regularities in the distribution of these tin ore deposits within each such Mesozoic metallogenic province are determined by the development of the tectonic and igneous activity of the given province.

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by

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• *translated by Eugene A. Alexandrov* •

ABSTRACT

Determination of the growth rate of crystals may be based on knowledge of the linear rate, the volume rate, and duration of crystallization. Direct measurements are possible where crystals are in process of growth as in the laboratory, saline waters, and at volcanic gaseous exhalations. Indirect but absolute measurements are possible by relating crystal zones with known occurrences in time; examples include calcite growth in belemnites, cave stalactites with smoke layers, and seasonally varied saline waters. Relative growth rates may be inferred from the nature of the intergrowth surface between different mineral crystals. Data is given on growth rates of gypsum, halite, calcite, fluorspar, quartz, hematite, pyrite, nepheline, bronzite, olivine, plagioclase, pyroxene, argentite, and corundum. -- M. Russell

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The rate of crystallization is a very substantial factor in the genesis of minerals. The size of separate mineral crystals and the amount of mineral formed depend mainly on the rate of crystallization. In spite of the importance of this factor, it has not yet been the subject of special consideration in literature. However, some ideas concerning the rate of crystallization must inevitably be considered during genetic studies of minerals.

Ideas concerning the dynamics of crystal growth are reflected by diagrams representing the succession of mineral formation. It should be said that these ideas are expressed clearly only by conventional terms in an appropriate division of physics. Perhaps this is considered in most detail by A. E. Fersman (1931, p. 672), whose recommendations are as follows:

"The time of deposition is represented on the diagram by a horizontal line. This line widens in places which correspond to the maximum of crystallization."

However, nothing has been told of how the process of crystallization was established and how it attained a maximum. Naturally, this property of mineral formation must be measured in some way. But in reality the lines on diagrams are plotted according to an evaluation by visual judgments, or impressions, which do not precisely reflect the amount or size of mineral units. It is difficult to draw conclusions on the apparent properties of minerals from a diagram of mineral formation sequence.

Meanwhile, data on the crystallization rate of minerals in nature may be obtained directly. While such cases are comparatively rare, they are extremely important, since they serve as the only basis for development of general concepts on the dynamics of mineral formation.

However, before examining data recorded in the field it is necessary to mention laboratory investigations. Numerous publications deal with experimental study of the dynamics of crystallization. One of the earliest is a paper by Kittl (1912), among the most recent are those of A. A. Leont'yeva (1943, 1947, 1948), who defined the linear rate of crystallization of pyroxenes, plagioclases, olivine, nepheline, leucite, magnetite, and hematite in melts. Such laboratory data are sometimes used in place of determinations in nature. It is understood, however, that experimental data are no substitute for natural observations, and can be used in mineralogy only for analysis of appropriate studies of natural minerals (Zavaritsky, 1943). We mention this information here to further correlation of experimental data with conclusions on the natural crystallization of minerals.

Let us use the following terms and symbols:

Linear Rate of Crystallization: (C)---linear accretion of mineral along the direction measured per unit of time.

Volume Rate of Crystallization: (V)---quantity (volume) of substance crystallized in a unit of volume per unit of time.

Duration of Crystallization: (T)---time interval during which the crystallization proceeds at one or the other rate.

Values sought may be measured directly during the process of mineral crystallization, but only when mineral formation is contemporaneous. In cases of past processes of mineral formation these values are deduced from products of crystallization.

Measurement of the rate of crystallization during contemporary processes of mineral formation may be performed in many cases: during crystallization of minerals in salt lakes and gulfs, in mineral springs, sometimes in the zone of oxidation of sulfide deposits, and during volcanic eruptions. It is regrettable that no such direct investigations are known in the field of mineralogy. In available publications only the fact of mineral growth during a specific T, or the results of growth which allow an approximate evaluation of the average rate of crystallization are recorded. Thus, C of halite along the normals to faces {100} in salt lakes, according to M. G. Valyashko (1952), equals a tenth of a millimeter per day.[3] According to an old communication by A. Breithaupt (1849), during the eruptions of Vesuvius in 1817 a fracture 1 meter wide ("more than 3 feet," p. 142) was filled with hematite during a period of 10 days. Hematite formed pneumatolytically by reaction: $2FeCl_3 + 3H_2O \rightarrow Fe_2O_3 + 6HCl$. The average C for hematite, without considering the direction of growth, amounts to 50 mm per day (by growth on both walls of the fracture). There are many similar data in the literature, but the C or V deduced from these data is very inaccurate. On the other hand, the role of contemporaneous processes in the genesis of minerals is very moderate.

Of greater interest is the possibility of determining the rate of crystallization during past processes of mineralization.

Determination of the absolute values of the crystallization rate in such cases is possible only when the minerals have, besides the indications of growth, some kind of time marks. These marks may be of different kinds.

First of all, time marks are represented by the zoning of individual crystals or by the banding of aggregates. The zoning and banding are the result of diurnal or seasonal changes of processes. According to Fersman (1952), crystals of gypsum which grow in the sediments of Sakskoye Lake in the Crimea have a zonal structure. This zoning is the result of inclusion of silt particles during seasonal changes in the growth of crystals and due to the introduction of silt by spring floods (fig. 1).

"After a study of these bands, " (i. e. zones of growth---D. G.) says Fersman (1952, pp. 812-813) "the following conclusions, supported by the facts of lake history, can be made: Growth of gypsum started about 1896. The crystal nuclei of gypsum formed in silt and grew

FIGURE 1. Rate of growth of gypsum in Sakskoye Lake, Crimea (approximately 2/3 of natural size.)

On the left - crystals of gypsum with growth bands, elongated in the direction of edge {111} : {111} .

On the right - interpretation of the chronology of crystal growth.

during the first 6 to 8 years within the mass of silt and mud. The normal and strong growth of regular bands (zones---D. G.) of gypsum began about 1902 after fairly abundant deposition of mud... There was a major break in the process of crystal growth in 1910-1912. The annual precipitation of gypsum was sharply reduced; for 3 to 5 years the crystals grew slowly. Finally, abundant deposition of gypsum in the form of yellow bands began during the last 2 to 3 years. "

In this example the T and accretion of crystals are known. The average annual C for gypsum from Sakskoye Lake is established according to these data along the edge (111) : (1̄11), along which, according to Fersman, the crystals are elongated. The linear rate of crystallization from the lake brine was 0. 4 mm per annum in 1900 and 3 mm per annum in 1903.

The study of gypsum indicates the rates of growth for quite recent processes.[4] The utilization of isotopic analysis allows us to find other time marks in the form of different seasonal

[3] According to M. P. Fiveg (1954) the average thickness of annual halite bands is 5 to 10 centimeters. This figure is also correct for sylvite. The annual bands of carnallite are 2 to 2.5 times as thick.

[4] The zonal distribution of inclusions due to seasonal changes is also observed in old formations, for example in halite of the Verkhnekamskoye salt deposit of Permian age (Dubinina, 1951). A very curious zoning of calcite stalactites is reported by Petranek and Pouba (1917) in Domica cave (Czechoslovakia) caused by soot from the fires made by ancient cave dwellers. This zoning indicates that the rate of stalactite growth is 0.5 to 2 mm per 100 years. L. Fisher (1934) collected data from nine localities in mines, dam tunnels, masonry of old forts, etc., in the United States where contemporaneous stalactites formed. According to Fisher the rate of growth is 0.33 to 7.5 cm per annum. It would be interesting to calculate from data of such kind the rate of growth of individual calcite crystals along different directions. In his pertinent article on the contemporaneous formation of calcareous stalactites, A. N. Churakov (1911) reported a rate up to 1 cm per annum.

absorption of various isotopes during very remote periods. Very interesting in this respect is the case described by H. Urey, H. Lowenstam, S. Epstein and C. McKinney (1954).

These authors determined the temperature regime of calcite crystal growth forming the rostrum of a belemnite which was collected probably in Oxford clay (Jurassic) at the island of Skye (Scotland). The temperature was determined according to the content of oxygen isotope O^{18}. They obtained an interesting graph which is here reproduced as Figure 2.
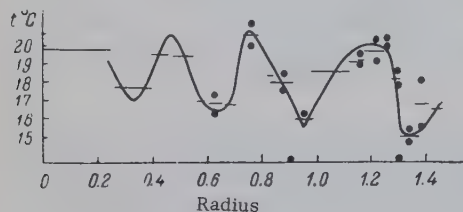
FIGURE 2. Temperature curve of belemnite growth. The temperatures are plotted in horizontal lines for the distance intervals, the average temperature of which they reflect. The dots indicate determinations on two opposite sides of the belemnite.

The rostrum has a diameter of 2.5 cm and clearly shows rings of growth in the cross section. It is regrettable that the authors do not indicate how the individual crystals of calcite are oriented with respect to the concentric zoning and the radiating aggregate. Studies conducted by the present writer on different samples of Jurassic and Cretaceous belemnites all have the same results: Calcite is oriented along the radii by its [0001] direction. We may therefore assume that calcite in belemnite from Skye Island grew along [0001].

The authors (Urey and others, 1954, p. 559) concluded that, in the belemnite they studied, three summers and four winters are recorded after the juvenile stage. The juvenile stage is reflected by a very small amount of carbonate insufficient to be studied by the present methods (p. 559). We find here in the form of seasonal temperature fluctuation the record of time marks which are necessary for our study.[5]

In this example, according to T and the accretion of crystals measured on the diagram of Figure 2 as distances along the abscissa, the average annual C of calcite in the direction of the maximum rate of growth along [0001] during

biogenic crystallization in belemnite is 0.31 mm for the first and second years of the "mature age" of the belemnite and 0.39 mm for the third year.

The zonal distribution of radioactive and radiogenic elements in the mineral gives us also a theoretical possibility of establishing the absolute age of chronologically different parts of the crystal. This is a third method of determining the rate of crystallization thus finding the value of T and accretion of zones.

Many analytical data are encountered in the literature on the zonal distribution of uranium, thorium and lead in uraninites from different deposits. However, judging by opinions on methods of investigation, such zonation cannot be considered in this sense, since such zoning is probably the result of mineral alteration from the surface.

Speaking about this third method of determining the absolute rate growth of minerals, we should say that the role of radioactive methods is still not clear. In evaluating methods of determining absolute values for the dynamics of crystallization during past processes of mineralization as a whole, one must stress their importance for our science and their value for the future.

Mineralogy faces and will continue to face the problem of formations containing no clues in this respect. However, in this case unique ways of investigation are opening, resulting in relative values of the rate of mineral crystallization.

The following method is based on correlation of the thicknesses of simultaneously formed bands or growth zones of different minerals. It is evident that thicknesses of such bands are proportional to the rates of mineral growth along the comparable directions. Under these circumstances the rate of growth of one mineral is considered conditionally as the measure of growth rate of the other mineral. This supplies relative magnitude of the value being determined.

The relative rate of growth in different parts of brookite from the northern Urals has been similarly determined when the parts have been depicted graphically (Grigor'yev, 1947).[6] Speaking of simultaneous crystallization of different minerals, one should refer to the fundamentals of the method.

Under conditions of joint crystallization of two different minerals, as well as of two different individuals of the same type of mineral,

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[5] For remarks on the authenticity of such a method of measuring the paleotemperatures, see Teys, Chupakhin and Naydin, 1957.

[6] The general aspects of this problem were examined by G. G. Lemmlein (1948).

surfaces of joint growth are created. These surfaces are represented by false faces. They possess the following features: The position and form of joint-growth surfaces and false faces relative to the elements of crystal facets depend on the orientation of crystals with respect to each other, and on the ratio of thicknesses of simultaneously growing layers of the adjacent faces. These surfaces are covered by induction striation running parallel to the intersection line of the adjacent faces of the adjoining crystals (Fersman, 1922; Frank-Kamenetsky, 1948; Shafranovsky and Grigor'yev, 1948; Lazarenkov, 1958).[7]

Figure 3 shows the three basic cases of evolution of mutual-growth surfaces between two different minerals, conformably with three examples. Mineral A started growing earlier at a uniform rate, while mineral B originated later and grew simultaneously on the surface of the first mineral. These cases are as follows: when the linear rate of crystallization of mineral B relative to that of mineral A is constant; when the rate of crystallization of
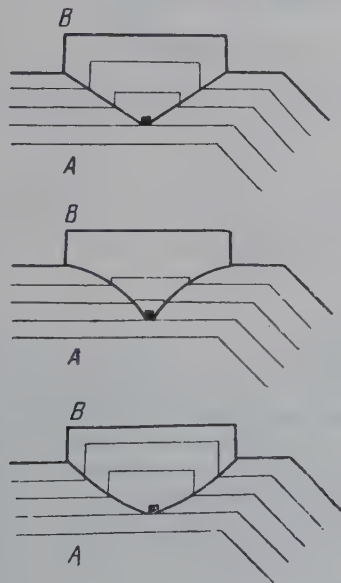
FIGURE 3. Three cases of formation of surfaces of joint growth between different minerals crystallizing simultaneously.

Top - the linear rate of crystallization (C) of mineral B relative to the C of mineral A is constant.
Middle - the C of mineral B increases.
Bottom - the C of mineral B decreases.

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[7] See remarks by the present writer relative to this feature, mentioned in the paper by N. Z. Yevzikova (1955, pp. 330-331)

B mineral increases, and when the rate of crystallization of B mineral decreases.

Quartz is a common mineral which crystallizes perfectly. It is the most suitable mineral for our study to determine the growth rate of other minerals.

In crystals from the Aldan basin quartz veins,[8] in the Kurumkan deposit for example, inclusions of fluorspar occur intergrown with quartz. Investigation of samples indicates that the quartz began to crystallize first and the precipitation of fluorspar began shortly before the quartz crystals stopped growing, but terminated before this moment. The octahedral fluorspar crystals are inclusions in the translucent quartz always at some distance beneath the surface. Since fluorspar occurs only at the top faces of quartz crystals, we must assume that its crystal nuclei were originally in solution and then settled on quartz, where they grew (Grigor'yev, 1951).

With more details the history of fluorspar formation becomes clear during examination of Figures 4 and 5.

The diagram in Figure 5 indicates that at moment 1 in the crystallization of quartz, marked by several closely spaced dusted layers, crystal nuclei of first-generation fluorspar settled on the surface of the upper crystal faces (fig. 5, Face $10\overline{1}0$). Some of these fluorspar embryonic crystals were located advantageously relative to the direction of maximum rate of growth G_4, and could continue to grow together with quartz. (There are two such crystals in fig. 5.) The others, which laid with face $\{111\}$ down on quartz were soon covered by growing layers of silica. Between the first and the second moments quartz and the octahedral fluorspar crystals formed simultaneously. Surfaces of joint growth, covered by induction striation, formed between these two minerals.

We note that during the individual fluctuations of the rate of growth of separate faces of fluorite crystals, the relation of crystallization of silica and fluorspar corresponded to the case represented at the bottom of Figure 3. At moment 2 the formation of fluorspar ceased, and quartz crystallized alone until moment 3, when fluorspar of the second generation appeared. Fluorspar of this generation formed new crystal nuclei and continued to grow on the top of protruding fluorspar crystals of the first generation. In this case, too, the possibility of growing developed only in crystals of fluorspar with advantageous orientation in the direction of highest rate of growth G_4. After a new time interval of joint growth (between 3 and 4),

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[8] These veins are described in a paper by E. M. Laz'ko (1957).

FIGURE 4. Fluorspar in quartz. The sample is oriented relative
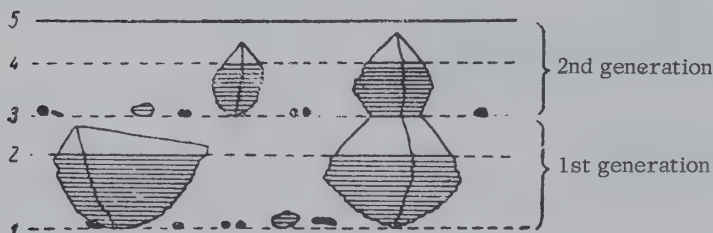to the direction of gravitational force. Kurumkan deposit,
Aldan. (Magnification 8)

FIGURE 5. Diagram of the joint growth of quartz and fluorspar.

The crystals of fluorspar are located on the upper face of
quartz crystal. The surfaces of joint growth of fluorite
and quartz are hatched. The figures 1-5 indicate the time
moments discussed in the text.

moment 4 is marked by a complete discontinuance of fluorspar formation, the crystals of which were next covered by quartz which formed until moment 5.

Let us determine the linear rate of crystallization of fluorspar and quartz. This is difficult to do for the entire process of joint crystallization because of the above-mentioned individual differences of C values for different faces of fluorspar crystals. Therefore we have to compare only the average values of C of both minerals.

Measurements of layer-growth thicknesses in quartz prism, which are visible due to cloudiness, and measurement of fluorspar crystal size, present the following values:

After the appropriate calculations when C of quartz along $\perp$ {$10\bar{1}0$} was recalculated for directions G_2 and G_3, and the last figure was considered as 1, C of quartz along G_2=0.18, along G_3=1; C of fluorspar along G_4=0.32.

It is also possible to make an attempt to

figure out the order of V value, which again will be relative and conditional. That is to say, it can be calculated if we evaluate the average amount of silica and fluorspar material deposited per unit of the quartz crystal surface. For this purpose we must determine the volume of the corresponding quartz layer on all of its crystal faces, though fluorspar occurs only on the top. Let us make a calculation for the period of crystallization of the first generation of fluorspar, assuming again the value of V for quartz equals 1. Then the values will be V of fluorspar=0.013, when V of quartz=1.

Figures of this kind probably reflect with some accuracy the relative amounts of mineral matter crystallizing from a volume unit of solution unknown to us during the time period under consideration. By the way, let us note that during the calculation we must determine the number of crystallization centers (N) of fluorspar. The average figures obtained (the distribution of fluorspar is irregular) for the upper faces of the quartz investigated are as follows: N of fluorspar per cm^2---first generation=9.8; second generation=4.8.

No succession in formation of centers of crystallization in time could be discerned (in contrast to experimental conditions).

In other quartz veins in the Kholodnoye deposit of the Aldan basin, quartz crystals contain inclusions of contemporaneous hematite. The crystallization of the latter occurred during the last stage of quartz formation, but terminated before the deposition of silica was finished. Therefore, hematite is covered everywhere by a layer of quartz.

In detail, the development of the process is as shown in Figure 6.
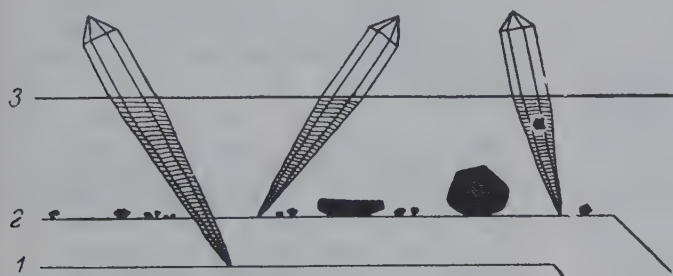
FIGURE 6. Diagram of joint growth of quartz and hematite. Kholodnoye deposit, Aldan (Magnification 3).

1, 2, 3 – zones of quartz growth (only the most important are indicated). Black – hematite – one large crystal is lying on quartz with the prism face {10$\bar{1}$0} ; another crystal is lying with the face {0001} . The bend of the induction surfaces of hematite is slightly exaggerated.

On first generation quartz crystals attaining considerable size at the time of moment 1, small quartz crystals of the second generation begin to grow from quartz fragments which fall from above. It is known that at moment 2 they are joined by quartz crystals of the third generation. In connection with this phenomenon let us mention that during the simultaneous crystallization of large and small individuals of the same mineral species, the larger crystals in the Aldan veins are growing ten times faster than the small ones. (Grigor'yev, 1949).

Crystallization of hematite, in which we are interested, begins near the formation time of third-generation quartz. Hematite has the form of hexagonal plates (a combination of the hexagonal prism {10$\bar{1}$0} and a pinacoid {0001}), which begin to split and change to the specularite, so characteristic of this mineral. Not all of the hematite crystals generated at the same moment continue to grow. The reason for this phenomenon cannot yet be indicated, but apparently it is not related to orientation. It should be noted that, beginning with the moment of hematite deposition, quartz takes on a smoky coloration.

During the simultaneous crystallization of iron oxide and silica, surfaces of joint growth form between quartz and hematite. These surfaces are visible on the crystals lying with their pinacoid face on quartz surface, as well as on crystals standing on edge. Because of a ratio of crystallization rates C different from that of fluorspar, the surfaces of joint growth form not a "cup, " but a flat "saucer". These surfaces are covered by the appropriate striation. The form of boundaries between hematite and quartz indicates that the crystallization of these minerals corresponds to the case shown at the bottom of Figure 3.

The measurement of relative values by conventional method results in the following figures: C of quartz along $G_2 = 0.3$, along $G_3 = 1$; C of hematite along $G_2 = 3$, and along $G_3 = 1.4$.

The method used above allows us to obtain relative values of the volume rate of crystallization. The result is that when V of quartz=1, V of hematite=0.15.

As may be seen from our second example with hematite, the rate of crystallization of a mineral correlated with quartz increases considerably as compared with the first example.

The average number of centers of crystallization N of hematite (nonuniformly distributed) on the upper faces of quartz was calculated for one of the samples studied. The N for hematite per square cm=7.2.

Having examined these two examples for the ratio of the rates of crystallization for two different minerals to that of quartz, let us now consider the crystallization ratio of quartz to the same mineral under different conditions. Such a mineral would be pyrite.

Scattered pyrite crystals, attaining 1 cm in size in our samples, occur in quartz veins of the Astaf'yevskoye deposit in the southern Urals (fig. 7).

In this case the crystal nuclei of pyrite generated on the upper face of quartz and continued to grow together with quartz. The crystals formed are a combination of the octahedron and the cube with very irregularly developed faces of the latter. As shown in Figure 7, one of the cube faces is much bigger than the rest of the faces of the same crystallographic form.

The crystallization of iron disulfide proceeded so fast relatively that the intergrowth surfaces of the pyrite acquired the form of a very flat saucer. The quartz layer deposited
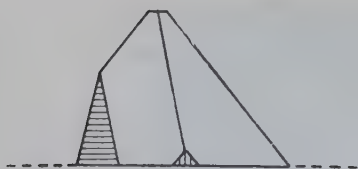
FIGURE 7. Drawing of a pyrite crystal in quartz. Astaf'yevskoye deposit, southern Urals (Magnification 2).

The octahedral form of pyrite is developed asymmetrically. The cube faces are hatched. The dashed line indicates the position of one of the quartz faces $\{10\overline{1}0\}$ on which pyrite grew. The bending of the induction surface of pyrite cannot be seen in such a small-scale diagram.

simultaneously with the growth of pyrite unmarked by any indicators (cloudiness or coloration), and its thickness is determined only by the character of surface bending of joint growth between the quartz and pyrite. As it should be, this surface is covered by induction striation, which is, however, very thin.

After the growth of pyrite suddenly stopped, quartz continued to crystallize and resulted in a thick cover over the pyrite.

The growth of quartz is determined in the direction $\perp \{10\overline{1}0\}$, according to the features recorded. To obtain the value of relative rate of growth of quartz along G_3 from analysis of facets, the ratio of crystal size along $\perp \{10\overline{1}0\}$ and G_3 has to be used in this case, since there are no visible indications of growth on the rhombohedron or on the prism. In samples this ratio averages about 4, but must be increased on account of the bottom parts of crystals which were broken off. Let us consider it as equal to 6, conditionally.[9]

Measurements and calculations indicate that C of quartz along $G_2 = 0.17$, along $G_3 = 1$, and C of pyrite along $\{100\} = 50$.

The approximate calculation of the relative value of V results, when V of quartz=1, in V of pyrite=1.15.

The number of centers of crystallization is very low in this case. N of pyrite per square cm=0.05.

Another example with pyrite is represented by a sample from one of the rock crystal deposits of the Pamir. Pyrite crystals of a complex form occur in quartz of this sample. According to macroscopical features this is apparently a combination of $\{100\}$, $\{110\}$, $\{111\}$ and $\{102\}$. At a given moment pyrite appeared here on a fairly large quartz crystal and proceeded to grow with it. It exhibits a perfectly formed surface of joint growth which is classified as "cup" type (bottom of fig. 3). Then the pyrite ceased to grow and quartz covered it with a thick layer.

The determination of relative C in this example is easily made for G_2 of quartz and $\{100\}$ of pyrite, but it is impossible to obtain directly the figure for G_3 of quartz, because the top of the crystal was broken off and no clearly visible zones of growth could be established on the rhombohedron. Assuming that the C value along G_3 is again 6 times higher than along G_2, we obtain the value of C along G_2 of quartz=0.12, along $G_3=1$, and C of pyrite along $\{100\} = 0.90$. For the same reason, the determination of V and N is here less certain, and therefore was not performed.

Let us now review and discuss all data we have obtained.

As is demonstrated by the determinations of the linear rate of crystallization of different minerals, the absolute values of C, their maxima, have a very wide range (table 1).

Comparing the absolute values of the linear rate of crystallization of different minerals

TABLE 1. The Absolute C Values of Minerals

Mineral	Direction	C (in mm per day)	Process of Crystallization
Gypsum	Edge $(111):(1\overline{1}1)$	0.01	Crystallization from brine in bottom deposits of muddy lake.
Calcite	[0001]	0.0011	Biogenic secretion from sea water in the body of belemnite.
Halite	G_4	0.n	From brine in salt lake.
Hematite	?	50	Crystallization in volcanic fracture during the interaction of gases.

during contemporary processes, we see that these values are 45,000 times that of the others (hematite — calcite). If instead of biogenic calcite, inorganic gypsum is taken into account, the factor is 4500.

We have data concerning only gypsum and

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[9] This relation could change several times during the process of crystallization.

calcite to establish the limits of fluctuation of C value for the same mineral. These data refer not only to crystallization of one substance under different conditions, but to the fluctuations of the value in which we are interested during different periods of time of the crystal growth observed. The greatest difference of the absolute values of the linear rate of crystallization obtained for gypsum is equal to 7.5.

The relative maximum values of the linear rate of crystallization of different minerals, as may be seen in Table 2, demonstrate also a wide range of variation.

TABLE 2. Relative C Values for Minerals When C for Quartz Along $G_3 = 1$

Mineral	Direction	C	Process of Crystallization
Fluorspar	G_4	0.32	
Hematite	G_2	3.0	In crystalline
Pyrite (Urals)	[100]	50.0	quartz veins.
Pyrite (Pamirs)	[100]	0.9	

In this case the range of C value is demonstrated by the values for fluorspar and pyrite from the Urals, the C value for the latter being 157 times that for fluorspar. Examples of crystallization of the same mineral, pyrite, in different veins demonstrate that C of pyrite in one vein may be 56 times that of C in another vein.

The volumetric rate of crystallization, represented conditionally, was determined only in relative values, the maxima of which are presented in Table 3.

TABLE 3. Conditional Values of V of Minerals When V of Quartz=1

Mineral	V	Process of Crystallization
Fluorspar	0.013	
Hematite	0.15	In crystalline
Pyrite (Urals)	1.15	quartz veins.
Pyrite (Pamirs)	0.025	

As with the linear rate, the V values differ for different minerals. The volumetric rate of crystallization of pyrite from the Urals is 88 times that of fluorspar. The factor 88 represents the range of the volumetric rate of crystallization. The same mineral, pyrite, in veins of the Urals demonstrates a V value which is 46 times that of pyrite from the Pamirs.

The values of N are also variable. But this is a special circumstance which is not presently being considered. (Grigor'yev, 1951).

The indicated variations of C and V depend on the individual crystallization capacities of the minerals themselves, on external geological factors, and, consequently, on the physico-chemical conditions of crystallization. The greatest ranges of absolute values are recorded for objects and processes remote from each other, such as the biogenic crystallization of calcite in a belemnite on the sea bottom, and formation of hematite from volcanic gases in a fracture. Lesser but still essential differences characterize the formation of different minerals under conditions of one type of process: In lacustrine deposits the absolute values of C for halite are 10 times the C for gypsum, while in quartz veins the C value for pyrite (Urals) is about 150 times that of fluorspar; the V value, 90 times that of fluorspar. It is especially interesting that the same mineral, pyrite, in veins of the same type demonstrates variations of C value of quartz which are almost 60 times — and V about 50 times — higher in one case than in another. Certainly, the latter data may be interpreted to mean that the rate of crystallization of the quartz itself is changed, but this means only a transposition in mineral name. It would be more correct to talk about considerable differences in the process of crystallization of the standard mineral as well as the mineral correlated with this standard from one deposit to another.

The rate of crystallization does not remain constant during mineral growth, and the above figures, as already defined, represent average values.

The character of alteration is not uniform.

Analyzing the annual process of halite crystallization in a salt lake, it is possible to conclude[10] that immediately after the start of crystallization there was during a springtime a gradual increase in C and V, a maximum of these values in summer, and a consecutive decrease of crystallization in fall and in winter, until the beginning of a new layer of salt the following spring. The seasonal factor evidently influenced the biogenic crystallization of calcite, without, however, interrupting the process.

Under endogenetic conditions in deposits of this type, the crystallization of fluorspar, hematite and pyrite together with quartz so developed that it started with a maximum relative value of C, followed by some slowing of the rate of growth, and, finally, a sudden and complete interruption of crystallization. An opposite case was described in the literature (fig. 8) when siderite as a companion of quartz demonstrated an increase of C, but crystallization suddenly

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[10] The deposition of salts in lakes may proceed due to the evaporation of solvent during the summer, due to cooling of brine in fall, and, finally, due to formation of an ice phase by the solvent in winter (Kashkarov, 1956).
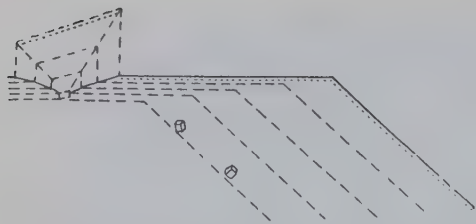
FIGURE 8. Diagram of joint growth of quartz and siderite. Northern Urals.

Top - quartz face $\{10\bar{1}0\}$, siderite crystallized simultaneously on its surface. On faces of quartz crystal which grew faster $\{10\bar{1}1\}$ the siderite is engulfed by quartz. The dots indicate siderite dusting.

stopped (Kornilov, 1954). If our examples were examined for quartz itself, they might indicate a gradual increase of C for this mineral, and accordingly a change of V values for all the minerals mentioned, being concluded again by a sudden interruption of crystallization.

An explanation for the peculiarities of mineral genesis in vein deposits should be sought in external factors. The problem is apparently as follows:

As has been established, the termination of crystallization in vein deposits of one kind of mineral and the beginning of the deposition of others, as well as a complete termination of mineralization, are usually caused in some way by tectonic movements. Specifically, these movements "command" the change of mineralizing solutions (by changing direction of flow). Depending on the physico-chemical factors, the process of crystallization begins, is disturbed, or is suddenly terminated by such mechanical phenomena.

In our examples the tectonics phenomena are clearly recorded by dust and "powder" deposited on the upper faces of the growing quartz crystals during repeated quakes. The beginning and end of crystallization of quartz and its companion minerals is associated with such moments. In the case of hematite in the veins of Aldan, there was also shattering. Fragments of quartz formed by shattering served as crystallization centers for the second and third generations of the mineral. It is difficult to tell whether the tectonics activity coincides with the beginning or the end of hematite crystallization (the interval is very short).

It is useful now to again turn to experimental data. They all indicate that the rate of crystallization of minerals is naturally far from the same order. According to E. Kittl (1912), who performed experiments with pure melts of a series of minerals, the maximum value of C for nepheline is greater than C for bronzite by a factor of 12. 5---naturally, for each mineral

at its temperature. At the same time, according to A. A. Leont'yeva (1947), the rate of joint crystallization of several minerals at one temperature level also differs. For example, in basalt melt (Kamchatka) the C of olivine at 1150°C is 0. 2 micron/min. , C of plagioclase is 1. 25 micron/min. , and C of pyroxene is 2. 25 micron/min. , it being known that the linear rate of crystallization of pyroxene is 11. 25 times greater than that of olivine (fig. 9). When the temperature is changed during the experiments the value of C also changes. The V values of minerals also vary as temperature changes.
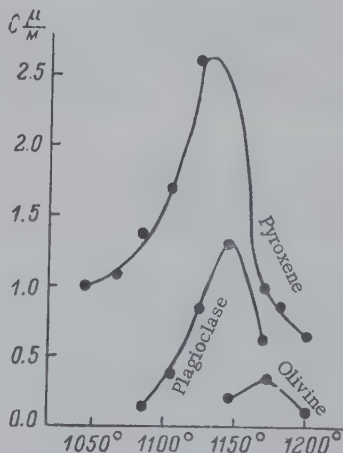
FIGURE 9. Temperature dependence of linear rate of crystallization of pyroxene, plagioclase and olivine in basalt melt. Kamchatka.

Let us present the following experimental data: There is an exothermic reaction during crystallization, while heat is absorbed during melting. A series of definitions of the heats of melting Δh, are involved in the latter process, expressed in joules per gram (Birch, Schairer, and Spicer, 1949, pp. 225-226). It becomes evident through calculation that when the appropriate minerals are removed from the melts, under temperature of crystallization and normal pressure, one calorie will be produced by crystallization of 0. 073 g of argentite, 0. 032 g of leucite, 0. 0189 g of fluorspar, 0. 00971 g of diopside, 0. 00947 g of anorthite, 0. 00389 g of corundum, etc. The difference in amount of forming crystals due to this factor is illustrated by argentite, which forms in a quantity 19. 3 times higher than corundum. If the solubility curves of minerals in water and other solvents are compared, different minerals again demonstrate a wide range of the amount of material crystallized under conditions of decreasing temperature per degree, or due to removal of a volume unit of solvent.

Summarizing this paper, it should be stressed that at the present time some possi-

bilities exist for determining by different methods both the absolute and the relative (linear or volumetric) rate of mineral crystallization. Using different methods, quantitative absolute and relative data on this important feature of mineral formation are found for a series of cases. Though the results are not yet numerous, and to a certain extent are of preliminary character, one cannot but admit their general importance. Data on minerals formed under conditions of free growth suggest a similar concept of crystallization under other conditions: metasomatism, recrystallization, etc. As may be seen, these results demonstrate a wide range of crystallization dynamics in natural processes. Without doubt, further studies will permit discovery of new peculiarities of this side of mineral genesis.

In conclusion, let us review those propositions on dynamics of crystallization depicted in the diagrams on the consecutive order of mineralization. In other words, let us note that the plotting of such diagrams is based usually on an assumed, but not formulated, yet quite clear statement about the equality of the volumetric and linear rate of crystallization of all minerals. Only on the basis of this statement is it possible to record the duration of crystallization from data on the quantity of minerals or the size of individual crystals or grains. This concept of crystallization features of minerals does not at all correspond to reality, and compromises at once the theoretical aspects of mineral genesis under consideration. We merely mention this here, since there is no sense in concentrating on other weak points of the criticized approaches. Nevertheless, perhaps it should be stressed that the utilization of the amount of mineral or size of individual crystals for determination of the duration of crystallization, generally speaking, is of dubious value. This becomes evident at once from a simple example. If a mineral crystallizes in the form of a cube with a constant C, then its V will change continuously with T according to parabolic law, and the volume of all accumulating substance will be represented by a cubic parabola. Conversely, if crystallization proceeds with a constant V, then C will change constantly, but according to an inverted parabola. Consequently, C and V in a general case are not at all mutually replaceable. (However, in some cases a change of V may be proportional to the alteration of C.) Besides, the value of V under condition of constant C is strongly influenced by the value of N. As a whole one would say that these widely known diagrams create only an illusion of knowledge, but are not of scientific value.

How should a diagram be plotted representing the genesis of minerals? It is understood that when the necessary data on dynamics of crystallization are lacking, only the consecutive order of mineral formation is recorded. The

problem of rate of crystallization remains to be solved during further investigations.

If data of such kind as are presented in this article are recorded for minerals which crystallized jointly, then only a relative duration T is established, as represented in Figure 10, for cases already described.[11] Concerning the quantitative representation of C and V on the diagram and their changing values during the course of the process, more detailed studies are required. We have had the opportunity of making only qualitative references to different cases of alteration in the relative values of C and V.
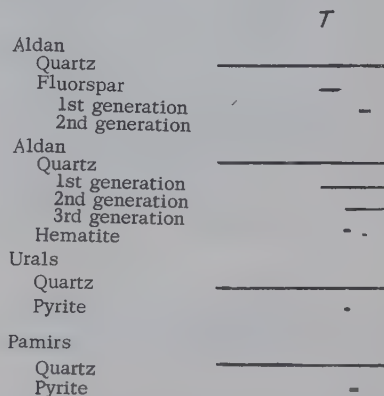
FIGURE 10. Diagram of the consecutive order of crystallization of minerals in quartz veins of the Aldan basin, the Pamirs, and the Urals.

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[11] See a similar diagram representing the above-mentioned case of crystallization of quartz with siderite in a paper by N. H. Kornilov (1954).

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THE POSSIBILITY OF DETERMINING TRUE TEMPERATURES
OF MINERAL-FORMING SOLUTIONS [1]

by

N. P. Yermakov and V.A. Kalyuzhnyy

• *translated by Vitaut Kipel* •

ABSTRACT

The development and application of thermometric analysis of liquid inclusions in minerals has reached an advance stage, but so many conflicting reports have been made that a state-of-its-art review is needed. Current capabilities and limitations of the method are outlined. --M. Russell.

* * *

About 1950, many possibilities were suggested for the use of inclusions of mineral-forming solutions in the study of the temperatures of crystallization and the physicochemical state of these solutions as related to mineral deposits.

At about the same time it became evident that only limited use, at best, could be made of minerals themselves for determination of relative and true temperatures of mineralization, and for distinction of products of pneumatolysis and hydrothermal processes. Since minerals themselves are of little use in such studies, many scientists turned to use of liquid inclusions as self-recording thermometers. Russian scientists were led to thermometric analysis and studies of the physicochemical state of ore-forming fluids. The geological and experimental investigation of the applicability of these methods in the determination of relative and minimum temperatures of formation of post-magmatic deposits have stimulated much discussion in the literature, both in Russia and abroad. This discussion has fostered understanding of the problem.

Studies of primary and presumably secondary inclusions confirm that changes of temperature can be recorded in different zones of single crystals, as well as in minerals of the same generation from varying depths and distances from the source of the solutions.

In many cases, the study of inclusions has permitted determination of relative temperatures of hydrothermal mineralization in coordinates of space and time, and to some extent determination of the limits of pneumatolitic and hydrothermal processes in polygenetic deposits.

The plotted results of thermal analysis of hydrothermal minerals illustrate either temporary stability, or frequent changes in temperature and related physicochemical parameters, permitting us to determine the dynamic history of hydrothermal processes in terms of relative data, such as temperatures of homogenization. Long ago, the temperatures of homogenization were considered as minimum temperatures of mineralization, rather than real temperatures. This caused dissatisfaction with the method among workers who though that determination of absolute temperature values was more important than understanding of the general range of temperature involved in mineralization. Some of these workers have suggested empirical corrections for temperatures at different pressures and concentrations of solutions (Ingerson, 1947, and Klevtsov, 1955). Some of these same scientists have stressed the errors of the method (Butuzov and Ikronikova, 1955, and Yakobova, 1955), forgetting the importance of determining relative temperatures of hydrothermal processes from temperatures of homogenization of liquid inclusions in crystals.

Recently, a series of articles in mineralogical journals discussed the reliability of the basic ideas of thermometric study of liquid inclusions, in the light of laboratory studies of natural material. Facts are sometimes considered in these articles without reviewing the entire process of mineralization. "New discoveries" being made are actually repetition of well-known facts about certain aspects of general behavior. This situation inspired the authors to publish this article. The purpose of this article is to consider the principles of the thermometric method in the light of modern data.

Liquid inclusions, which are minute natural "autoclaves", were studied to determine the basic factors in the process of formation: relative and absolute temperatures, pressure, concentration, chemical composition, and physicochemical state of solutions.

Since in the last five years doubt has been expressed about the thermometry of mineral-forming solutions, the authors will discuss

only these questions, even though this is not the most important aspect of the study of inclusions. The authors believe that at present it is possible to make a preliminary survey of recent discussions, and to try to explain some of the premature conclusions which often arise from incidental or accidental interest in the study of inclusions.

Minerals as temperature indicators in hydrothermal processes at best provide only minimum and maximum values; only the inclusions of liquid aqueous solutions offer any hope of understanding of the dynamics of the process.

The study of homogenization of such inclusions attempts to duplicate, in reverse order, the formation of the inclusion and to restore the original pressure-temperature-concentration parameters. Temperature is considered to be the leading parameter because it can be easily changed without disrupting the system as a whole. Changes of temperature in a closed system cause concomitant changes in the related variables, as shown by the PT diagram (figure 1). During heating, the moment of disappearance of the gas bubble in the two-phase inclusion corresponds to the lower point of the liquid-vapor equilibrium curve. Thereafter, the state of the system, nonhomogeneous, will be characterized by the coordinates of points on a corresponding curve (isochore) representing equal specific volumes. The temperature of homogenization is considered a minimum value for formation of the inclusion, because the isochores are not parallel and form an acute angle with the pressure axis.

Pressure corrections based on PT diagrams of systems of corresponding solution concentrations will permit a closer approach to the true temperature of mineral formation (Ingerson, 1947). PT diagrams for pure water can be used as a first approximation.

Temperature error due to dissolved host-mineral in solution at formation will be insignificant if this concentration is not high. This concentration is established by chemical analysis of the inclusion or examination of the substance deposited on the walls of the vacuole (Lemmleyn and Kliya, 1952). Study of the solubility of minerals in solutions resembling liquid inclusions will allow closer correlation of homogenization and true formation temperatures.

Let us consider the effect of components such as easily soluble salts and gases, often present within the inclusion, on the precision of temperature measurements. The presence of such compounds makes use of the PT diagrams for water imprecise. In such systems, PT diagrams are required for each value of concentration. The first important experimental attempt to establish a PT relation curve for the system H_2O - 30 percent NaCl was made by P. V. Klevtsov (1955).

A high concentration of salts within the inclusion widens the limits of application of the homogenization method, raising the critical temperature of the solution. Volatile compo-
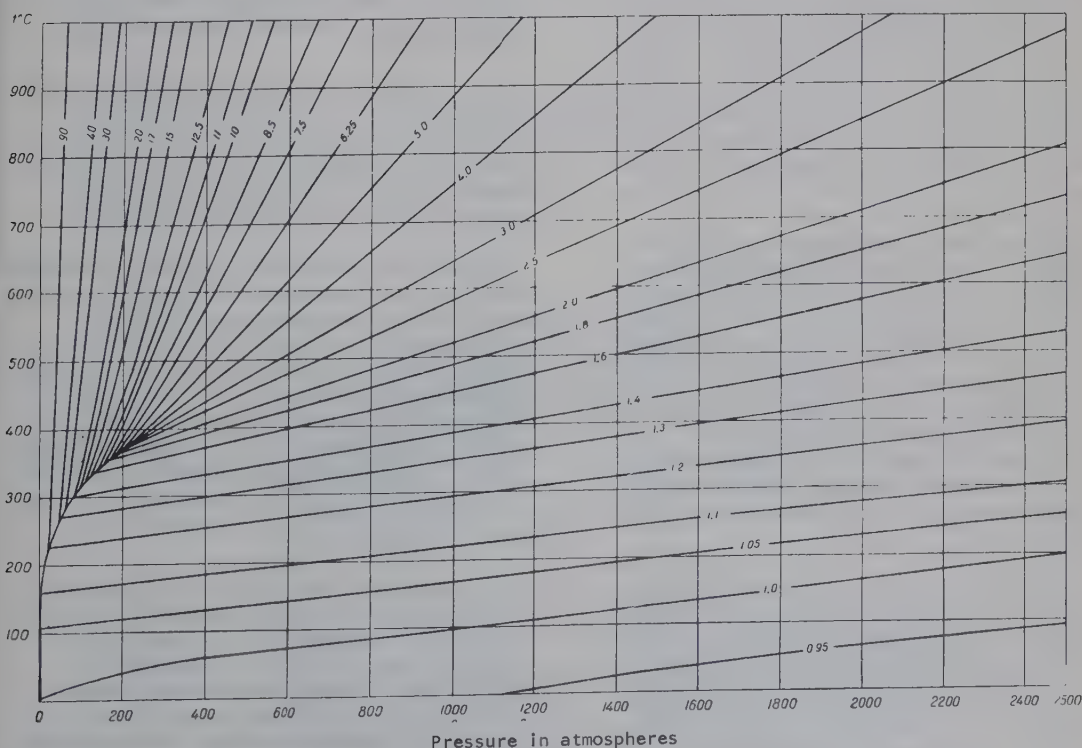
FIGURE 1. Isovolumetric diagram for water according to Kennedy and others

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nents (such as CO_2, H_2S, etc.) influence the accuracy of thermometric measurements in a another way, by lowering the critical point of the solution, thereby reducing the limits of application of the method. Inclusions which contain carbonic acid, as well as gaseous inclusions, are not recommended for use in thermometric analysis. Equilibrium curves for such solutions show deviations toward higher pressures than do curves for pure water. The angle between isochores of inclusions and the pressure axis will increase with increasing volatile content, while homogenization temperatures will deviate more from real temperatures of mineralization (see figure 1).

Less important factors cause differences between homogenization temperatures in particular cases:

The preservation of inclusions in a mineral usually occurs in a homogeneous environment. In our opinion, formation of a multiphase inclusion can occur only in relatively rare cases, such as: 1) preservation of gas bubbles adhering to growing crystal faces during solution boiling; 2) during solution boiling and separation of volatile component (for example, separation of CO_2 due to decrease of CO_2 solubility in aqueous solution); 3) due to adhesion of microcrystals of associated minerals to the faces of a growing mineral, and formation of vacuoles with inclusions nearby.

It should be stressed that the conditions of crystallization with closure of the heterogeneous system are easily recognized by detailed study of the crystal. The rare case of determination of inclusions formed by enclosure of gas bubbles from a boiling solution (Yermakov, 1950) presents some difficulties. (The boiling is limited by the critical pressure of the given solution.)

Temperatures determined from study of "impaired" inclusions [The impairment is caused by the deposition of mineral matter on the walls - V. K.] cannot generally be used, even for determination of minimum formation temperatures. The mechanism of formation of such associations of inclusions was studied for the first time by Lemmleyn and Kliya (1952) during experiments with soluble salts. The phenomenon of impairment of inclusions surely occurs in natural minerals, particularly due to a rapid temperature decline at the end of the process.

Inclusions which have been considerably altered during cooling can be readily detected, and are not suitable for these studies. Long experience in these studies indicates that most minerals contain many more inclusions with constant phase relationships than impaired inclusions. The former are sometimes found in the order of several thousand per cubic

centimeter. Contrary to the opinion of some authors, impaired inclusions are exceptional, rather than normal. V. V. Yakubova (1955) stresses impaired inclusions, providing eleven photomicrographs and drawings of samples with abnormal phase relationships, and no samples of inclusions with normal behavior. The selection of samples of inclusions in minerals from Murzinka by Yakubova was not logical and paid no attention to the papers of Lemmleyn, Kliya, and others. Neglecting distinction of primary and secondary inclusions, she obtained uncertain and contradictory data. For example, according to studies based on obviously secondary liquid inclusions and complex inclusions with carbon dioxide in pyrogenetic ichthyoglypts of quartz, Yakubova (1955, pp. 133-134) attempted to establish the temperatures of formation of quartz. Here she demonstrated that her knowledge of the subject may be compared with the knowledge of Neptunists. It is known that in α-quartz of ichthyoglypts the only primary inclusions are solidified inclusions of residual melt, and that the wide range of gaseous and liquid inclusions illustrate a long process of influence of later solutions acting on graphic pegmatites at varying pressures and temperatures.

Significant errors are introduced in this problem by the lack of PT diagrams for a wide range of concentrations of aqueous solutions. PT diagrams for water, useful for a close approximation of actual solution behavior, were first used in these studies by Nacken in 1921. A detailed study of the subject was done by Ingerson in 1947.

Recent papers on the subject by Ostrovsky (1950) and Vulchin (1951) use VTX and PTX diagrams to make important conclusions about heterogeneous equilibria in binary systems at constant volume. This application filled a hiatus in the theory concerning this method.

While these conclusions are correct relative to inclusions in easily soluble minerals, thermometric analysis is limited in application to slightly soluble minerals. Besides the basic parameters on the VTX diagrams, many details of secondary importance are reflected which illustrate the process of phase equilibria in a broader aspect. Many of the elements of these diagrams are insufficiently supported by factual data, especially in the case of the system SiO_2-H_2O. Some of the details can obscure the essence of the problem.

Vulchin (1951), basing his work on that of Ostrovsky, mistakenly reaches the conclusion that ordinary homogenization temperatures should be higher than actual mineralization temperature. Careful study shows the values of homogenization and true temperatures quite close at low pressure, with the homogenization temperature too low at higher pressures.

Errors corresponding to those suggested by Vulchin would occur in inclusions formed in boiling solution, when the mineral can capture part of the heterogeneous system. Impaired inclusions and inclusions containing CO_2 would also produce such errors. Such situations can be readily identified, and would not account for major errors. Because boiling of mineral-forming solutions is limited by critical pressure, formation of inclusions of heterogeneous origin is not a widespread phenomenon (except those containing CO2), especially in medium and low temperature hydrothermal deposits.

Recently, some authors (Butuzov and Ikornikova, 1955), discussing the possibilities of application of thermometry in mineralogy, made erroneous assertions which disagree with the theoretical concepts of heterogeneous equilibria, namely with the basic PT diagrams of thermometry.

Some scientists working with synthetic quartz crystals were surprised to find homogenization temperatures were constant for crystals produced at constant concentration, constant degree of a autoclave filling, and varying temperatures. Had the authors introduced corrections for pressure, the true temperatures of formation would have been determined. Butuzov and Ikronikova, unaware of the need for the correction, concluded wrongly "the temperature of homogenization of gas-liquid inclusions is not related to the temperature of crystal growth". Had they noted that the state of material inside the autoclave was, in

this case, characterized by the same isochore on the PT diagram, this incorrect conclusion could have been avoided. As we see to our regret, these authors failed to see that the relationship between temperature of homogenization and temperature of crystallization is certainly indirect. If this relation were direct, all of our reasonings about corrections and physicochemical diagrams would be unnecessary. The binding link between the two values is the specific volume of the solution. Determining the temperature of homogenization, we obtain the specific value of solution at the moment of enclosure, or we find on the PT diagram a corresponding and entirely different isochore. Determination of absolute temperature of past processes of mineralization requires knowledge of the pressures involved. Having determined the corresponding isochore on the PT diagram by measuring homogenization temperature, we can say that at the pressure (determined or estimated independently) only one temperature value could exist. We now have an equation with two unknowns. When the pressure at a constant constant concentration is known, such an equation can be solved.

Considering the pressure corrections recorded during recent experiments, probable pressures during postmagmatic processes should be mentioned.

High pressures do not seem to be typical for the hydrothermal process, which usually takes
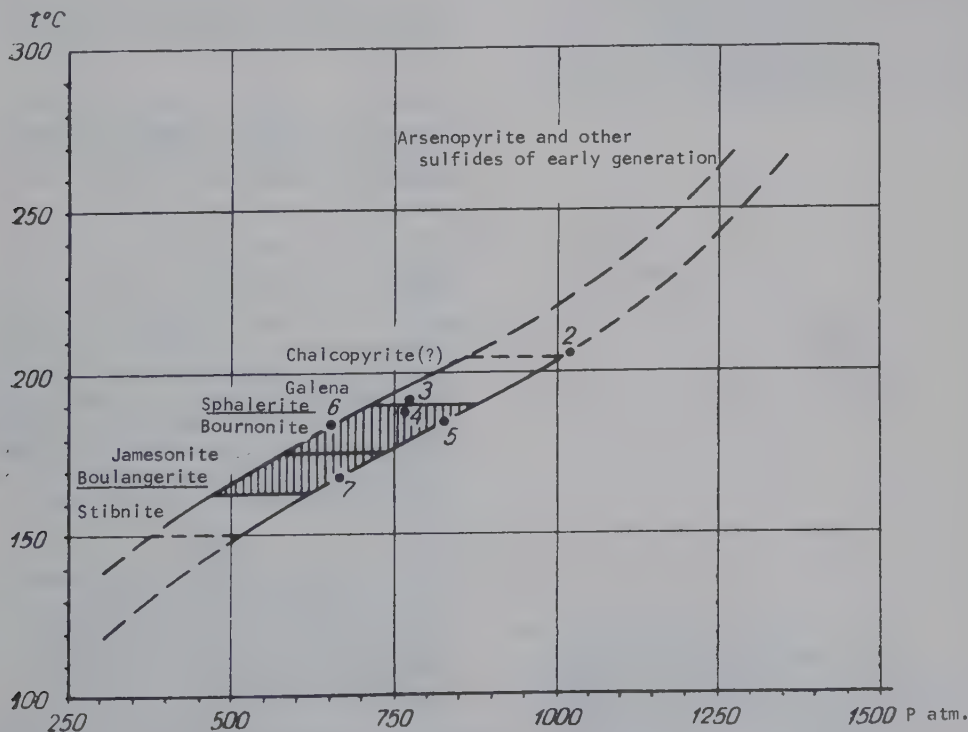
FIGURE 2. The field of deposition of ore minerals in Nagol'ny Kryazh in coordinates T-P. Figures represent numbers of the experiments

place in open channels of faults and fractures. Pressure in these channels is not lithostatic, but rather hydrostatic, increasing 100 atmospheres per kilometer of depth. It should also be stressed that the system is far from closed because of the permeability of even undisturbed igneous rocks. Hydrothermal and pneumatolytic systems in particular are usually open systems. The high pressures of autoclaves are not common in nature except in the magma chamber.

A few determinations of pressure using inclusions containing CO_2 or brine solutions indicate pressures of 300-500 atmospheres for pegmatites and less than 820 atmospheres for gold and base metal hydrothermal deposits. Pressure as well as temperature changes continuously in time and space, making limits hard to define. Pressures seem usually to be below 1,000 atmospheres. (see fig. 2).

Solutions at low pressures and concentrations of 12 to 15 percent investigated at the VNIIP Institute revealed new data of homogenization temperatures in synthetic quartz crystals prepared by hydrothermal synthesis. Results of these investigations were presented at the present symposium by Safronov and Khadzhi (1957, table on p.57, [of orig. journ.,ref. footnote 1.--M.R.]

For pressures within the range from 230 to 600 atmospheres and true temperatures of 316° to 330°, the temperatures of homogeniza tion of inclusions and their deviation from the true temperatures were determined. As a result of these experiments an empirical curve of pressure corrections (Δ degrees) was plotted for the temperatures of homogenization of two-phase inclusions (figure 3).
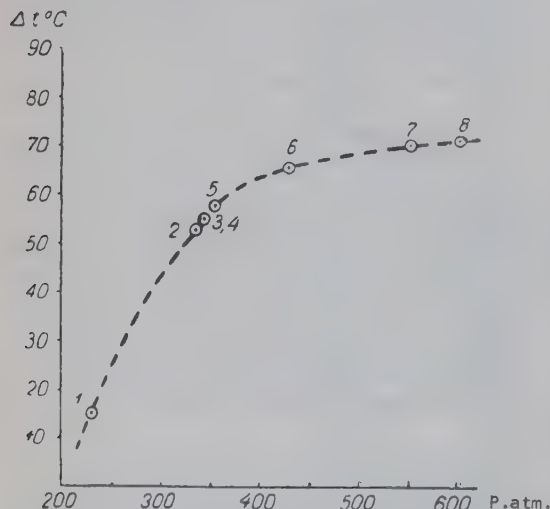
FIGURE 3. Pressure corrections (Δ t^o degrees) of temperatures of homogenization of liquid inclusions according to data recorded during growing of synthetic quartz crystals in solutions of 12 to 15 percent concentration. The figures indicate numbers of the experiments

The corrections varied from 15° to 71°. This difference and the correction diagram (figure 3) are related to homogenization temperatures obtained in conventional air-heated bombs with imperfect sealing. In sealed vessels with contact heating of a sample between metal plates, corrections will average 40° lower; and at 600 atmospheres, corrections will not exceed 30-31° (except experiment no. 1).

For pressures up to 1,600 atmospheres and 30 percent NaCl concentrations (corresponding to polyphase liquid inclusions), Klevtsov (1955) established empirical temperature corrections of 30 to 40^OC. at 1,500 atmospheres where homogenization temperatures were in the range of from 170 to 400^OC.

Thus, using the existing methods of study of inclusions for determining values of density, concentrations, and pressures in solutions by means of thermometric analysis (Kalyuzhny, 1955; Smith, 1953; Yermakov, 1950), one may substantially come closer to the determination of true temperatures of the natural process of mineral formation.

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STUDY OF ERUPTIONS AND EARTHQUAKES
ORIGINATING FROM VOLCANOES (PART 1 OF 3)[1]

STATISTICAL RELATIONS BETWEEN ERUPTIONS AND EARTHQUAKES OF ASAMA VOLCANO

by

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•   *translated by Kinkiti Musya*   •

ABSTRACT

Reports of volcanic eruptions and earthquakes originating from volcanoes indicate that seismic activity preceding the eruption is related not only to eruption magnitude and structure of the volcano, but also to viscosity of the lava at the time of eruption. This follows, since lava of higher viscosity meets greater resistance as it ascends from the magma chamber to the earth's surface and, consequently, greater stress will be produced within and beneath the volcano. The writer gives a condensed statistical breakdown of earthquakes and explosive eruptions of Asama Volcano. The Asama earthquakes treated in the report are mainly those of rather low magnitude (T = 1.0 sec, V = 350) at the Asama Volcano Observatory, situated 4.2 km east of the center of the summit crater. This investigation showed that most of the explosive eruptions were preceded by an increase in micro-earthquakes. In addition, an experimental formula for predicting volcanic eruptions, based on the statistical relation between frequency of earthquakes originating from Asama and its explosive eruptions. The forthcoming report (Part II) will discuss the same problem based on seismic observations by more sensitive instruments set nearer the summit crater. -- A. Eustus

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PREMONITORY PHENOMENA OF VOLCANIC ERUPTIONS

Prediction of volcanic eruptions to prevent disasters may have been considered at an early stage in human history.

In essence, prediction of volcanic eruptions amounts to understanding premonitory phenomena of eruptions and determining the relation between these phenomena and subsequent eruptions. The following premonitory phenomena of eruptions have been noticed in many volcanoes: 1) Changes of earthquake frequency in the vicinity of the volcanoes, 2) crustal deformations including upheaval, subsidence, tilting, and expansion and contraction of the earth near volcanoes, 3) changes of geothermal gradient, 4) changes of earth's magnetic field near volcanoes, and 5) changes in the nature and the volume of emitted steam and gases.

Of these, 1) and 2) are dynamic phenomena. Sometimes both occur simultaneously and these phenomena are more closely interrelated than any of the other phenomena. Considering their genetic interrelation, this is quite reasonable. In this report, mainly problems concerning 1) will be considered.

EXAMPLES OF EARTHQUAKES PRECEDING VOLCANIC ERUPTIONS

It is well known that six eruptions of Usu-san (Omori, 1911; Minakami et al., 1951) in historic times (1663, 1768, 1822, 1855, 1911, and 1944-1945) were preceded by noticeable earthquakes. Frequent earthquakes and topographical changes were particularly noticeable up to 6 months prior to the first of the 1944-1945 eruptions, and have been reported in detail. Topographic changes that occurred at the eastern foot of the volcano were most notable. Upheaval amounted to over 100 m and a lava dome was formed on the upheaval area. Lava erupted in this case was highly viscous dacite (10^{11}-10^{13} C.G.S.) (Murase, 1958).

Earthquakes occurring prior to the great Martinique eruption of 1902 have been reported by Lacroix (1904). In this case, a spine of highly viscous lava was protruded through the steep slope near the summit of Mount Pelée. Subsequent collapse of the lava spine caused great havoc in the city of Saint Pierre and other communities at the foot of the volcano.

Prior to the great 1914 eruption of Sakura-jima (Omori, 1916) local strong earthquakes occurred frequently, some strong enough to damage buildings in Kagoshima City. It is noteworthy that many small craters formed in an approximately straight line extending from east to west on the flank of the volcano, centered on Minami-dake crater. Great quantities of lava were also emitted during the 1941 eruption, but, according to the seismic observations with a Wiechert seismograph at the Kagoshima Weather Station, (1946) no noticeable earthquake occurred, although the eruption was preceded by the increase of microseismic activity. One reason why the mode of occurrence of pre-

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monitory earthquakes in the 1941 eruption differed from that in the 1914 eruption is thought to be that the later eruption occurred along a pre-existing line of weakness.

The relation between eruptions and premonitory earthquakes of Asama Volcano, which ejects andesitic lava similar to Sakura-jima, will be described later in more detail. Explosive eruptions have occurred frequently in Asama in recent years, but it is noteworthy that few large earthquakes have taken place during the last 40 years. However, microseisms have occurred up to ten times in a single day, even when the volcano was quiescent. Several hundred microseisms have occurred up to ten times in a single day, even when the volcano was quiescent. Several hundred microseisms occurred in a single day prior to explosive eruptions during periods of activity. In short, explosive eruptions of Asama Volcano are preceded by earthquakes, but because these earthquakes are small in magnitude and shallow in origin, they cannot be recorded unless seismic observation network of rather high magnification is established and properly arranged.

In addition, the relation between premonitory earthquakes and volcanic eruptions which spewed out basaltic lava such as Kilauea (Macdonald and Wentworth, 1952; Macdonald et al., no date; Furumoto, 1957), Mauna Loa, or Mihara-yama (Takahashi et al., 1939; Kizawa, 1951; Tazawa, 1957) will be described. According to a report of the Kilauea Volcanological Observatory, a local earthquake strong enough to inflict considerable damage on the island of Hawaii occurs once every several years. Many noticeable earthquakes originate from depths of several to several tens of kilometers, but the relation between eruptions and these earthquakes is not clear. On the other hand, in Miyake-shima, Mihara-yama (in O-shima), and in the vicinity of the volcanic islands, noticeable earthquakes occur in swarms from time to time. Though the eruptions of Miyake-shima in 1940 (Tsuya et al, 1941) and of Mihara-yama in 1950-51 were rather large and were accompanied by the extrusion of lava, large earthquakes regarded as forerunners of eruptions were not recorded. Actually there is almost no recorded example of eruptions from basaltic volcanoes being preceded by premonitory earthquakes. However, there are very few volcanoes where observation has been continuously carried out with high magnification seismographs. This might be one of the reasons why premonitory earthquakes have not been recorded.

The following conclusion can be drawn on character of earthquakes preceding volcanic eruptions and the nature of lavas ejected by them.

I. Eruption of highly viscous lava is preceded by large earthquakes. Eruption of andesitic lava sometimes is accompanied by large earthquakes and sometimes by microseisms (in a few volcanoes where seismometric observa-

tions have been made). Eruptions of fluid basaltic lava are almost without exception not accompanied by quakes.

II. The characteristics of pre-eruption earthquakes vary considerably, even if the erupted lava is the same. First is a case when an open vent connects the crater and the lava reservoir, and lava is easily erupted. Second is the case of a volcano where eruptive activity had ceased for many years and the volcanic vent is plugged with solidified magma. In this case, lava is forced through a newly opened conduit by the rising column of magma. As a result, pre-eruption earthquakes are much more numerous than in the case where the volcanic conduit was free from obstruction.

In short, the difference between I and II may be related to the degree of resistance which the lava encounters in rising from the lava reservoir to the surface of the earth.

DYNAMIC PRE-ERUPTION PHENOMENA AND VISCOSITY OF LAVA

Characteristics of earthquakes, crustal deformations, and changes of the earth's magnetic field differ in quantity and nature according to the type of volcanoes involved, and are related particularly to the physical and chemical properties of the ejected lava. It is evident from the above-mentioned examples of volcanic eruptions that dynamic phenomena are closely related to the viscosity of ejected lava. That is, generally speaking, the viscosity of andesitic or dacitic lava (Hagiwara et al., 1946) is 10^5 - $> 10^{10}$ poises at the temperature of $1,000^o$-$1,100^o$ C. at the time of eruption. In such cases pre-eruption phenomena such as earthquakes and crustal deformations are noteworthy. On the other hand, the temperature of basaltic lava at the time of ejection is, in many cases, on the order of $1,100^o$-$1,200^o$ C. and the viscosity (Minakami, 1951) is 10^2 - 10^5 poises. Thus, the viscosity of basaltic lava is considerably lower than andesitic or dacitic lava. In the case of basaltic eruptions, pre-eruption dynamic phenomena generally are not noteworthy. Considering a simple model, the increase of pressure (P) in the lava reservoir when magma (viscosity coefficient: η and density of magma: ρ) rises and is ejected at the mean velocity (v) through a vent (radius: a, length: l) from the magma chamber is represented by

$$P_1 = \frac{8\eta v l}{a^2} + \frac{1}{2}\rho v^2$$

and when magma rises through a dike (width : 2b, depth :l) and feeds a fissure eruption,

$$P_2 = \frac{3\eta v l}{b^2} + \frac{1}{2}\rho v^2$$

As shown in the above formula, if other con-

ditions remain constant, the increase of pressure which causes eruptions is a function of viscosity. Moreover, the viscosity of dacitic lava is more than 10^7 times the viscosity of basaltic lava. Consequently, considering only this fact, it is evident that dacitic or andesitic eruptions produce great stresses inside and under the volcanoes. As a result, remarkable earthquakes and topographical changes occur. These are the pre-eruption phenomena of the volcanic outbursts.

The foregoing is a very simple model and all ramifications of eruption phenomena are not included. For example, in volcanoes which have been dormant for many years preexistent dikes and vents are plugged with solidified lava. Eruptions do not occur unless volcanic bodies are destroyed or open fissures are re-established. The increase of stress in the lava reservoir must become great enough to destroy part of volcanic superstructure. Earthquakes and topographical changes occur during destruction of this superstructure. The writer cannot but consider that the differences between the dynamic pre-eruption phenomena discussed above, are closely related to variation of internal stress which is a function of magma viscosity. On the other hand, basalt exhibits magnetization several times stronger than andesite or dacite. Moreover, the changes of magnetization due to the changes of temperature in the interior of volcanic bodies are remarkable, and sometimes the changes are observed as the changes in the earth's magnetic field. For this reason, pre-eruption phenomena of basaltic eruptions can be detected from geomagnetic observation. This is illustrated by remarkable changes of the earth's magnetic field in Miharayama (Rikitaki, 1951; Yokoyama, 1957; Nagata, 1941; Minakami, 1941; Takahashi and Hirano, 1941) and Miyake-shima.

EARTHQUAKES ORIGINATING FROM VOLCANOES

As described in the preceding paragraph, the mode of occurrence of earthquakes varies with the nature of the ejected lava and with the facility or difficulty of eruption (i. e. with the existence or nonexistence of an open volcanic conduit). The case of Asama-yama will be considered from this viewpoint. Firstly, the lava viscosity is intermediate between that of basalt and dacite. Concerning the structure, in many cases lava of high temperature is exposed in the bottom of the summit crater, indicating the existence of a conduit connecting the crater with the deep interior of the volcano. Therefore, it is apparent that the resistance against rising lava is not unduly great, but the resistance is greater than Kilauea and Mihara-yama from which highly fluid lava is discharged. Therefore it is probable that Asama-yama receives greater stress during periods of eruption than do comparable basaltic volcanoes.

Earthquakes originating from Asama volcano are classified into four types, on the basis of their epicenters and the relation between earthquakes and eruptions.

Type 1: A-type. These are earthquakes originating from a rather deep (1 to several km.) and extensive area under the volcano. Earthquakes of this kind are very few in number and only two or three have been observed during the last 40 years. However, Omori has reported that many earthquakes were felt during the remarkable 1911-1914 activity. For convenience, the writer names these A-type earthquakes.

Type 2: B-type. These are earthquakes originating in swarms from and near the bottom of the summit crater. Their foci are very shallow (within 1 km.). Most earthquakes originating from Asama Volcano belong to this category. None of these earthquakes are felt by local inhabitants, and most are microearthquakes less than 10μ in amplitude, measured at the station several hundred meters from the epicenters. These earthquakes are characterized by an indistinct S phase and the predominance of surface waves. The writer calls these B-type earthquakes.

Type 3: Explosion earthquakes. These are earthquakes accompanying explosive eruptions. The strength of earthquakes of this type is proportional to the strength of eruptions. Generally speaking, the motion of these earthquakes resembles that of B-type earthquakes, but the amplitude is larger and sometimes reaches more than several hundred μ at an observation point 4 km from the crater. However, noticeable earthquakes are rarely observed because the earthquakes motion has a large vibration period. The writer calls earthquakes of this kind explosion earthquakes.

Type 4: Volcanic microseisms. Continuous vibrations of microseismic type. Vibrations sometimes appear when minor eruptions continue for a long time or when fumarolic activity in the crater bottom is unusually great. The writer calls vibrations of this kind volcanic microseisms. Volcanic microseisms correspond to harmonic tremors defined by G. A. Macdonald and others (no date) in a report on Hawaiian volcanoes.

It has been reported by the present writer that earthquake types 1 through 4 were observed during the activity of Minami-dake (South Peak) of Sakura-jima in recent years. Earthquakes originating from Asama-yama and Sakura-jima in general resemble each other in nature and mode of occurrence, but there are some slight differences between earthquakes originating from the above two volcanoes. In the case of Volcano Sakura-jima, earthquakes of A-type predominate in frequency and magnitude; volcanic microseisms also are noteworthy. However, volcanic microseisms originating

from these two volcanoes are not comparable with those of Mihara-yama, Aso-san, and the Hawaiian volcanoes.

For example, with regard to earthquakes originating from the Mihara-yama area, A-type earthquakes (1) occur relatively frequently and sometimes in swarms. In Mihara Volcano, earthquakes corresponding to B-type earthquakes of Asama-yama and Sakura-jima occur as microseisms rather than as single earthquakes originating from a very shallow depth in the crater area. Explosion earthquakes also cause continuous vibration when accompanying Strombolian eruption in which small explosions occur at intervals of several seconds to more than ten seconds. In addition, the amplitude of volcanic microseisms accompanying continuous eruptions is usually larger than that of volcanic microseisms usually associated with B-type earthquakes.

The 1943-1945 activity of Usu-san, when Shōwa-Shinzan was constructed, was very remarkable. Earthquakes of A-type and B-type which occurred during this activity were noteworthy, but microseismic earthquakes were unimportant.

The types of earthquakes originating from various volcanoes were described in the foregoing paragraphs. Regarding these types of earthquakes, it is apparent that the nature of the earthquakes is related to the type of volcanic eruptions and ejected lava, i.e., that the physical property of the ejected lavas is related to the nature of eruptions which in turn governs the nature of the resulting earthquakes.

ERUPTIONS OF ASAMA VOLCANO AND EARTHQUAKES (B-TYPE)

Earthquake foci which occur in Asama volcano will be described later in some detail. As mentioned above, most earthquakes which occur in that volcano are B-type which originate in swarms from and near the crater bottom. So the relation between the eruptions of Asama volcano and the frequency of earthquakes is merely the relation between eruptions of the volcano and the frequency of B-type earthquakes.

In order to clarify this problem, the relation between eruptive activities of Asama volcano from 1934 to 1952 and the associated volcanic earthquakes which originated from the volcano is shown in Figure 1. Here, volcanic earthquakes are the 10-day sum of the number of earthquakes observed with a low-magnification seismograph (magnification: 350, period: 1 sec.) at a station 4. 2 km east of the crater. Explosive eruptions are the one-month sum of kinetic energy of the solid ejecta. Comparison and examination of the above two phenomena reveals the following:

1) Seismic and eruptive activities are almost coincident in time.

2) When seismic activity is great, eruptive
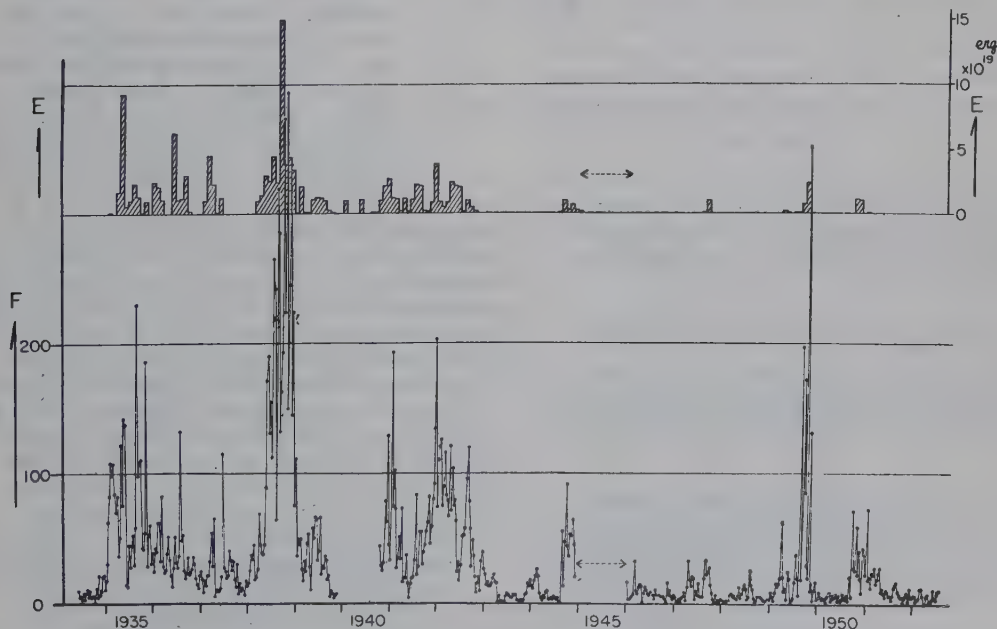
FIGURE 1. Seismic activity (F) originating from Mt. Asama and its explosive eruption (E).

activity also is great, and when the frequency of earthquakes is low, no eruption occurs.

3) Closer examination shows that in many cases the frequency of earthquakes increases prior to eruptions. Because Figure 1 is of such small scale, the relation between the seismic activity and the eruptive activity of Asama volcano will be shown later in another way.

The parallel relation between earthquake frequency and eruptive activity in Asama volcano is better seen in Figure 2, which shows the sum of the annual frequency of earthquakes versus the annual energy of explosive eruptions. According to Figure 2, more than 200 earthquakes were observed in a year when no eruption occurred (with the above-mentioned seismograph 4. 2 km east of the crater).
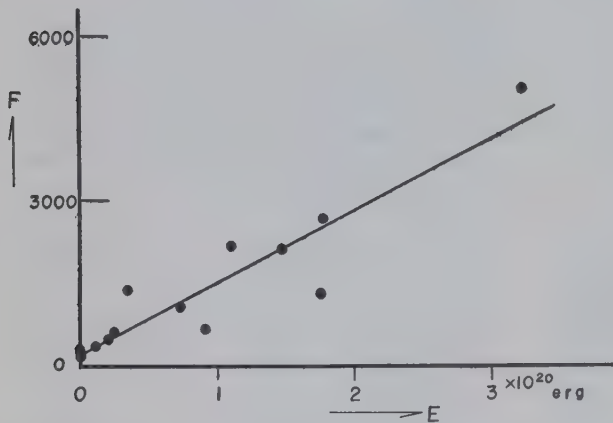
FIGURE 2. Relation between annual frequency of earthquakes originating from Mt. Asama and annual energy of explosive eruption of the same volcano (1935 1950)

The relation between the number of earthquakes (annual frequency) F and the energy of eruptions (annual energy) E is represented by the following formula:

$$F = 1318E + 223 \text{ (Unit of E: } 10^{20} \text{ ergs)}$$

Undoubtedly the frequency of B-type earthquakes and explosive eruptions are related because most of the points plot approximately on a straight line.

In many andesitic volcanoes beside Asama, the rise and decline of eruptive and seismic activities are approximately parallel, if the two phenomena are compared, disregarding a phase difference of 2 or 3 months.

OCCURRENCES OF EXPLOSIVE ERUPTIONS AND EARTHQUAKES

The relation between explosive eruptions and

earthquakes which was described in the foregoing paragraph will be discussed in some detail. For this purpose the relation between an individual explosive eruption and the frequency of earthquakes was studied. It was discovered that earthquakes in many cases increase in number before, during, and after individual eruptions. Close attention was paid to the frequency of earthquakes preceding eruptions.

I. Sometimes one to three weeks before explosive eruptions the frequency of earthquake swarms reaches a maximum. Figure 3 shows the frequency of earthquakes preceeding the eruptions of December 4, 1941 and of September 3 and 21, 1949.

II. Sometimes the occurrence of earthquakes increases from one to two weeks preceding explosive eruptions. Figure 4 shows the occurrence of earthquakes which preceded the explosive eruptions of November 7, 1935 and March 25, 1938.

III. On occasion the frequency of earthquakes is highly variable, and sometimes conspicuous pre-eruption earthquakes do not occur. The earthquakes which preceded the explosive eruptions on April 20, 1935, May 24, 1939, and February 27, 1937, are examples (fig. 5).

As described above, most eruptions are preceded by an increase of earthquakes, though, quantitatively speaking, a particular eruption and its earthquake have no definite relation. Sometimes eruptions take place without a notable increase of earthquakes, although such cases are rare. This conclusion is the result of observation with a low-magnification seismograph at a station 4. 2 km distant from the crater. Consequently the number of earthquakes observed is several orders of magnitude fewer than the number of earthquakes observed with a high-magnification seismograph set up near the crater, which is the epicenter of these earthquakes. Thus, it is very rare that an increase of earthquake frequency, as compared with a quiescent period, does not precede the occurrence of eruptions. However, there is no difference, quantitatively speaking, in the respect that eruptions and earthquakes do not always appear in the same mode. The results obtained with the high-magnification seismograph will be discussed later.

SOME STATISTICAL RESULTS ON THE FREQUENCIES OF ERUPTIONS AND EARTHQUAKES

It is necessary to investigate statistically the nature of earthquakes originating from volcanoes and the relation of these earthquakes to erup-
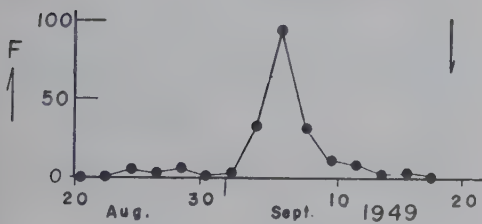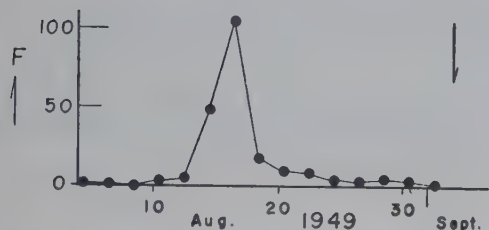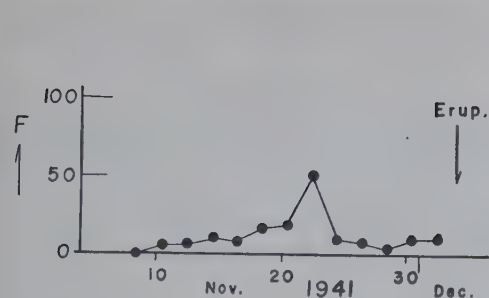
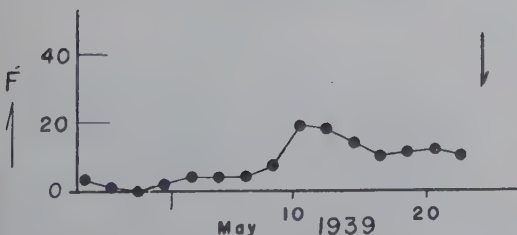FIGURE 3. The development of seismic frequency preceding to respective eruption (pattern I).

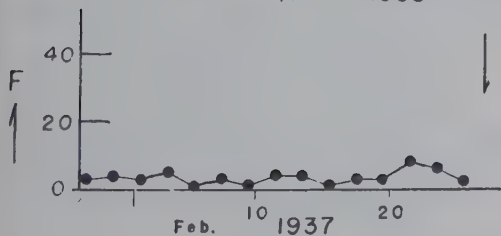FIGURE 4. The development of seismic frequency preceding to respective eruption (pattern II).
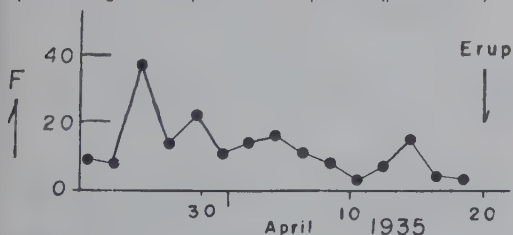

FIGURE 5. The development of seismic frequency preceding to respective eruption (pattern III).
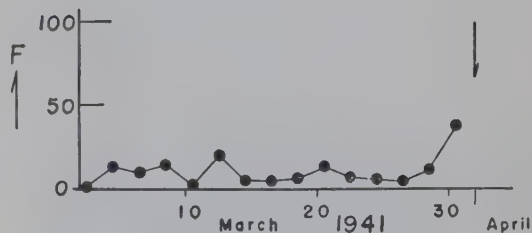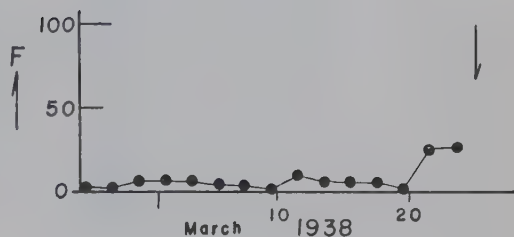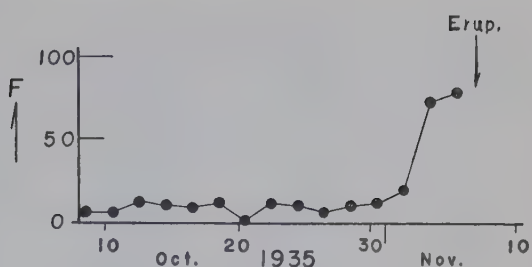
tions. Attention is focused on the observations of Asama Volcano from 1934 to 1952 (excluding 1945); these observations are shown in Figure 1. The frequency of earthquake occurrences will be investigated in some detail.

I. In order to clarify the daily frequency of earthquakes originating from Asama volcano, the frequency distribution (F_0) of the daily frequency (n) of earthquakes in the above period was investigated. As the mean daily frequency of this distribution is 2.339 and the standard deviation (σ) of the daily frequency is 11.68, the above frequency distribution is in accordance with the Polya-Eggenberger's distribution (F_c) represented by the above value, i.e., a distribution of the type in which earthquakes take place in swarms. The observed and calculated values of n are shown in Table 1.

II. Next, taking every 10-day sum (N) of the frequency of earthquakes, the relation between an individual eruption and the frequency of earthquakes was examined. Or, obtaining the distribution of N in every 10-day sum of the daily frequency n_1, n_2, n_3, ...

$$n_1 + n_2 + n_3 + \cdots\cdots\cdots + n_{10} = N_{10}$$
$$n_2 + n_3 + n_4 + \cdots\cdots\cdots + n_{11} = N_{11}$$
$$n_m + n_{m+1} + n_{m+2} \cdots n_{m+9} = N_{m+9}$$

Table 1.

Number of earthquakes	Frequency distribution	
	Observed value	Calculated value
0	1972	2097.0
1	1137	982.4
2	693	622.9
3	419	429.3
4	257	307.8
5	208	225.8
6	175	168.1
7	110	126.5
8	78	95.9
9	60	73.2
10	57	56.1
11	48	43.2
12	36	33.4
13	34	25.8
14	14	20.0
15	19	15.6
16	18	12.1
17	11	9.5
18	10	7.4
19	10	5.8
20	12	4.5
21	7	3.5
22	4	2.8

The curve (F) in Figure 6 represents the frequency distribution throughout the above period obtained by dividing the frequency into 0-14, 15-29, ... On the other hand, the curve (F') represents the frequency distribution of N in the previous day of eruptions (that is N'). The mean value of N through the above period is 32. 4 and σ^2 is 109. 6. The mean value of N for the days prior to eruptions is about twice the mean value of N throughout the period. Hence, it is evident that the occurrence of earthquakes is remarkable for 10 days prior to eruptions. Moreover, the ratio of the frequency distribution throughout the period represented by the curve (F) to the curve (F'), using the value of N as the parameter, is serviceable as a basic datum for predicting an explosive eruption on the following day.
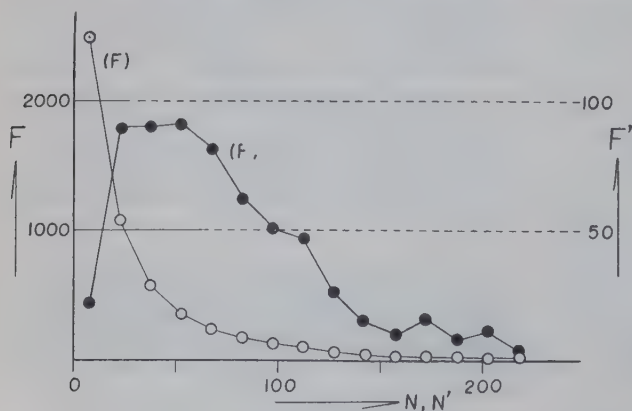
FIGURE 6. The frequency distribution (F) of N for the period from 1934 to 1952 and that (F') for the previous days of explosive eruptions which occurred in the same period.

III. In II the frequency distribution was examined by dividing the values of N into equal intervals such as 0-14, 15-29, ... This time the frequency of the intervals 0, 1-2, 3-5, 6-9, 10-14, ... was extended with the values of N, and the frequency distribution throughout the period (F) and the frequency distribution in the previous day of explosive eruptions (F) was examined in the same way as a method described in II. The relation between the values of N and the vulnerability ratio of explosive eruptions calculated based on F'/F is shown in Figure 7. According to Figure 7, when the value of N is less than 5, explosive eruptions rarely occur on the following day. When the value is more than 30, the probability of an eruption on the following day is more than 0. 1, and when the value exceeds 100, the probability is about 0. 5. In short, Figure 7 shows that as earthquakes become more frequent, the likelihood of an eruption increases.
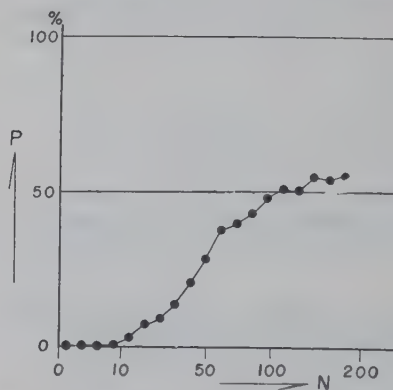
FIGURE 7. Probability distribution useful for predicting explosive eruption of Volcano Asama, based on seismic frequency for ten days.

Examples showing the relation between explosive eruptions and the frequency of pre-eruption earthquakes were shown in the previous paragraph. If the total frequency of earthquakes is taken for only 20 or 30 days before eruptions, no marked difference from the above result is recognized statistically. However, when the vulnerability ratio of eruption based on earthquake frequency for various days is calculated and put to practical use, it is hoped that it will prove useful for predicting volcanic eruptions.

CONCLUSION

The effectiveness of predicting volcanic eruption based on the statistical study of previous eruptions will depend on whether or not the results can be applied statistically to future eruptions. The above-described statistical results are believed to be completely satisfactory for the mode of occurrence of explosive eruptions and earthquakes in

the future.   This problem will be discussed in detail in a forthcoming report.

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[Parts 2 and 3 will appear in subsequent issues of International Geology Review].

Reference Section

This section is devoted to a listing of selected geologic items appearing in the two publications of the Library of Congress; Monthly Index of Russian Accessions, and East European Accessions Index. These lists are intended as a means of indicating to researchers in the earth sciences some of the material most recently available for screening, further review, and translation. For this reason the lists do not include material now, or soon to be, published in English. Emphasis is placed on Russian material; the extent to which items from East European sources are listed depends on the country and language involved.

A major function of the AGI translations program is the screening of foreign literature for material that should be made available to the English-speaking scientist. Researchers who need such material are urged to review these lists and send us their recommendations for consideration by the editors; the translation needs of all geologists will be served better thereby.

-- Managing Editor

MONTHLY INDEX OF RUSSIAN ACCESSIONS

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Lower Cretaceous stratigraphy of the northwest-ern extremity of the Greater Caucasus. V. L. Egoian. pp. 14-33.
Remains of siliceous organisms in middle Maikop sediments in the Northern Caucasus and their correlative value. A. K. Bogdanovich. pp. 34-37.
Origin of fossil medusae. V. A. Grossgeim. pp. 38-41.
Petrographic studies using the immersion method. V. A. Grossgeim. pp. 42-49.
Typomorphic characteristics of Paleogene sedi-ments in the western Kuban and their im-portance for correlation and paleogeography. V. T. Malyshek. pp. 50-53.
Zeolite in Sarmatian sediments of western Kuban. V. T. Malyshek. pp. 54-56.
Bajocian and Bathonian stages in the Kuban-Belaya interfluve. K. O. Rostovtsev. pp. 57-66.
Oil and gas potentials of Miocene sediments in Krasnodar Territory. I. P. Zhabrev. A. A. Dvortsova. pp. 67-74.
New data on Maeotic sediments in western Kuban. V. N. Buriak. pp. 75-78.
Geology, and oil and gas potentials of northern Fergana. G. M. Aladatov. pp. 79-89.
Chemical composition and zoning of underground waters in the Azov-Kuban oil- and gas-bear-ing region. V. S. Kotov. pp. 90-96.
Change in the composition of natural gases under the effect of the underground oxidation of hydrocarbons. V. E. Narizhnaia. pp. 97-112.

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Using data of borehole radiometry in estimating reservoir properties of strata when searching for sites suitable for underground storage of gas in the Kaluga region. V. V. Larionov. pp. 3-15.
Relation between the porosity parameter, the co-efficient of surface conductivity, the diffusion-

REFERENCE SECTION

RECENT TRANSLATIONS IN GEOLOGY

A review of the Translation Services

This part of the Reference Section is devoted each month to a listing of the new translations of geologic significance which have become available from sources other than IGR and the established cover-to-cover journals in geology. This is done to accomplish several purposes: 1) inform geologists of the foreign literature in their field available in translation; 2) provide information necessary to avoid duplication of translation effort, and 3) advise geologists of the activities of the various organizations providing translations or related services in their field.

AGI TRANSLATIONS POOL

The list of recent translations in geology carries this month the first translations to have been submitted in response to an announcement in GeoTimes (v. 5, no. 8, p. 14) of the establishment of an AGI Translations Pool. Translations sent in will be indexed, publicized and made available to interested scientists at cost of photocopying. The intent is to make as widely available as possible, at the lowest cost to researchers, translations which have not heretofore been placed in public repositories.

GEOLOGIC-TRANSLATION JOURNALS

The following translation journals regularly contain material pertinent to geology. The subsequent list of recent translations does not include articles from these journals:

Atomic Energy, published by Consultants Bureau.

Bulletin (Izvestiya) of the Academy of Sciences U.S.S.R., Geophysics, published by the American Geophysical Union.

Doklady of the Academy of Sciences of the U.S.S.R., Earth Sciences Section, (Geochemistry, geology, geophysics, hydrogeology, mineralogy, paleontology, petrography, lithology and permafrost), published by the American Geological Institute.

Geochemistry, published by the Goechemical Society.

Geodesy and Cartography, published by the American Geophysical Union.

Izvestiya of the Academy of Sciences of the U.S.S.R., Geologic Series, published by the American Geological Institute.

Petroleum Geology, published by the Review of Russian Geology.

Problems of the North, published by the National Research Council of Canada.

Soil Science, published by the American Institute of Biological Sciences.

Soviet Geography, selected translations and reviews published by the American Geographic Society.

Soviet Physics: Crystallography, published by the American Institute of Physics.

SOURCES OF TRANSLATIONS

Lists of recent translations from which the following list was selected are:

Technical Translations, vol. 5, nos. 8, 9, 10 and 11.

NLL Translations Bulletin, vol. 3, nos. 6 and 7.

An index of sources and addresses will be found at the end of the list of translations.

Geologists and translators are invited to submit titles which have not been cited by services from which we compile these lists. The submittal of a copy of the translation itself will be construed as an offer for IGR to publish, make copies available at cost of reproduction and/or consign it to a major translations repository at our discretion. Suggestions for improving this service are welcome.

RECENT TRANSLATIONS

Anonymous, 1960, Geology, exploration, weather forecasting and cartography (selected articles): trans. from Jen Min Jih Pao (Chinese People's Republic), 15 Dec. p. 7; 18 Dec., pp. 4 and 7; 19 Dec., p. 7. JPRS: 4435 OTS 61-21305. $0.50.

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